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THE UNIVERSITY OF ALBERTA

SNOWMELT INFILTRATION AND SEASONAL GROUNDWATER INTERACTIONS WITH THE UNSATURATED ZONE.

BY

© Charles Peter Maulé

A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH IN PARTIAL FULFILMENT OF THE REQUIREMENTS FOR THE DEGREE OF DOCTOR OF PHILOSOPHY

IN

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Charles Maulé

583 St Patrick St
Victoria, British Columbia
Canada V8S 4X4

DATE: August 12, 1989

THE UNIVERSITY OF ALBERTA
FACULTY OF GRADUATE STUDIES AND RESEARCH

The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research for acceptance, a thesis entitled Snowmelt Infiltration and Seasonal Groundwater Interactions with the Unsaturated Zone submitted by Charles Peter Maulé in partial fulfilment of the requirements for the degree of Doctor of Philosophy in Soil Physics.

D.S. Chanasyk (Supervisor)

S. Pawluk

R. Rothwell

K. Muehlenbachs

H. Steppuhn (External)

Date: Aug 16, 1989

To my parents, Peter and Lindsay,

for all I am learning from them.

ABSTRACT

Since snow accounts for up to 30% of the annual precipitation on the prairies, it may serve as a source of soil moisture for crop growth. The influence of management practices, particularly tillage, has not been extensively researched. Overwinter and springmelt soil moisture dynamics were investigated for several tillage treatments; standing, chiselled, and disced stubble; and summer fallow at Ellerslie, Alberta from fall 1985 to summer 1987. Soil moisture was measured with a neutron probe, while contributions to the root zone and the groundwater (at 4 m) were estimated with the use of the environmental isotopes, deuterium and oxygen-18. A numerical flow model tested explored the hypothesis of soil water flow which seasonally alternated flow directions.

The chiselled and stubble treatments had significantly greater snowmelt recharge than the disced and summerfallow. Summerfallow had the smallest snowmelt recharge. Standing stubble resulted in the most soil water at spring seeding. Although wetter fall conditions resulted in smaller snowmelt recharge to all treatments, the order of greatest to least recharge was unchanged.

Although summer precipitation differed from winter precipitation by 12‰ $\delta^{18}O$ units, a snowmelt wetting front could not be discerned. This was likely due to soil conditions which resulted in mixing of waters rather than displacement and to the larger than normal variation of isotope content in summer rains. Based upon the seasonal differences of isotope concentrations in precipitation, soil water and groundwater were calculated to be composed of 11 to 19% and 44% snow waters, respectively.

Groundwater was 56% composed of water that did not reside in the root zone long enough to be altered by evaporation. This implies the occurrence of fast draining waters from near saturated surface events.

Soil moisture and water table measurements indicated seasonal soil water flow with upward flow occurring during the winter to a freezing front and downward flow in the spring through fall seasons. Numerical simulation using alternating flow resulted in matching simulated isotope contents with observed. Soil moisture flows are likely less than 0.08 cm/d and upward flow exceeds downward flow by 1.5 to 2 times.

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The thesis is merely the end of one particular journey, although at times it has seemed long. Upon reflection of this journey, I have found that I appreciate the journey itself - in consideration of the goal, the thesis. Surprisingly it is not just the specifics within this thesis that I have gained, but that of a certain appreciation and approach to the study of science, and also that of friendships. For this I extend thanks to the people in the Departments of Soil Science and Geology and the Alberta Research Council.

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1. INTRODUCTION

1.1 BACKGROUND AND OBJECTIVES

The Canadian Prairies may be generally defined as having a subhumid to semiarid climate; the potential evapotranspiration is greater than the precipitation which ranges from 300 to 500 mm per year (Longley 1972). The term 'prairies' refers to a large treeless tract of level or undulating grassland. This region comprises the southern parts of the provinces of Manitoba, Saskatchewan, and Alberta. Soils in well drained sites belong to the Chernozemic Order. The first impression of explorers and early settlers of this region now known as the Canadian Prairies was that it was a desert, incapable of supporting viable crop production. This concept delayed the introduction of farming to both the Canadian Prairies and the Great Plains of the USA (Cannell and Dregne 1983). To European settlers or to those from Central Canada, the Prairies certainly seemed like a desert compared to their more humid climes. The fertility of Prairie soils and their ability to hold moisture eventually dispelled this notion; but occasional droughts, especially without moisture and soil conservation procedures, have brought back these first impressions.

The Prairie climate is characterized by long dry periods, resulting in a moderate to severe soil moisture deficit, so that "supplementary irrigation and moisture conservation practices can be of value in almost all agricultural areas in almost all years " (Laycock 1960). To counter this deficit and to achieve maximum crop production, an additional 30 to 50 mm of soil water in stubble fields is required for much of the Prairies (PFRA 1982). Summerfallowing has been the mainstay for providing this additional moisture but it has also resulted in increased erosion and salinization. As snow accounts for 20% to more than 30% of the annual precipitation

(Steppuhn 1980), it can be considered a viable alternative source of extra soil moisture to fallow; and if properly managed can provide opportunity for continuous cropping practices (Chanasyk 1986).

The climate of the Canadian Prairies is distinctly seasonal. The seasons of winter, spring, summer, and autumn may be defined as four different periods of three calendar months each beginning with December 1, March 1, June 1, and September 1 respectively (Longley 1972). Winter if characterized by monthly temperatures below 0°C or/and a permanent snow cover may last five months, November through March. Although snowmelt generally begins in the last week of March, spring is considered to be from 1 April until the end of May (Longley 1972). While the winter months are the driest the summer months, June, July, and August generally provide 40% of the annual precipitation. Fall thus lasts from September 1 to November 1, when a permanent snow cover usually develops. Temperatures range from -20°C during the winter to +30°C during the summer.

Studies of soil water during the winter have been few, due to the impression that little of significance is happening during this period; vegetative growth ceases, the soils are often frozen to depths surpassing 2 m (Maclean 1974), and it is the driest time of the year. Investigation proves otherwise. The root zone (surface 50 to 100 cm) and shallow water tables (within 3 to 5 m) respond to seasonal fluctuations of temperature and precipitation. Upward water migration from the water table to the freezing front can occur (Willis et al. 1964; Benz et al. 1968) resulting in a significant contribution to soil moisture under certain conditions (Gray et al. 1983). Also, dissolved salts may be carried by fluid movement toward the freezing front (van Schaik and Rapp 1970; Cary and Mayland 1972; Landsberg 1981).

Mid-winter snow depth are common on the prairies, resulting in partial infiltration and refreezing of the melt waters, which reduces soil permeability during springmelt. Meltwater does not penetrate the soil deeply until the ground has thawed, which does not occur until most of the snow cover has melted (Murray and Gillies 1971; Granger et al. 1977; Gray et al. 1983). Thus much of the accumulated snow can occur as runoff. Runoff from stubble and tilled fields commonly varies between 50% and 90% of the snow water equivalent (Willis et al. 1961; Willis et al. 1969; Patterson 1985). Snowmelt runoff may accumulate in local depressions resulting in springtime 'groundwater mounds' in which the local water table rises above the 'normal' regional water table (Meyboom 1966). By summer the standing water has disappeared due to infiltration and evaporation. Transpiration demands of vegetation in and around the depressions can result in a reversal of water movement, from that during spring ponding, in which water flows from the water table to the near surface zone (Meyboom 1966).

To date most of the snowmelt infiltration studies have used soil moisture measurements (Gray et al. 1983; Chanasyk 1986) and/or infiltration measurements (Kane and Stein 1983), but few have used tracers to investigate the movement of the melt waters once they enter the soil (Saxena 1984; Bengtsson et al. 1987). The use of tracers could provide information about: the mixing and displacement of soil waters by the infiltrating melt waters, the contribution of snowmelt to groundwaters, and the seasonal interaction of groundwaters with soil waters. If the environmental isotopes deuterium and oxygen-18, which are naturally present in meteoric waters, are used then several advantages over other tracers may be realized. These advantages include: a characteristic seasonal fluctuation in abundance, uniform application across the

landscape with meteoric waters, provision of 'double' labels enabling evaporated waters to be discerned, and non-reactive tracers. It is the first feature, seasonal fluctuation in abundance, that makes these isotopes most attractive for use tracers of snowmelt. Edmonton, Alberta had the second largest variation in seasonal isotopes amongst 93 global IAEA sites (Yurtsever and Gat 1981). The difference in average oxygen-18 abundances ($\delta^{18}O$) between summer and winter precipitation is 12 ‰ (IAEA 1969, 1970, 1971). As oxygen-18 may be measured to ± 0.05 ‰ (Gonfiantini 1981), this large seasonal difference makes environmental isotopes attractive as tracers in soil and groundwater studies.

Understandably much emphasis on soil water movement and soil moisture conservation practices have been given to the growing season. However, during the five months of frozen winter conditions of the Canadian Prairies, processes still continue that will have an effect on the soil moisture regime of the growing season. The objective of this study is to improve the understanding of the effect of Canadian winter conditions upon the soil moisture status of agricultural lands of the Prairies. Of concern to this objective is the overwinter snow accumulation, the contribution of snowmelt to soil water and groundwaters, and the seasonal interaction of the unsaturated and the saturated zones.

Results and discussion of the study are presented in five chapters in this dissertation. All chapters are written as independent papers, each with its own specific objectives. These objectives are summarized as follows:

1. *To determine the effect of fall cultivation practices upon snowmelt infiltration.* Snow depth and soil moisture content were monitored over two winter periods. The effects of fall cultivation upon snowmelt recharge were evaluated through measured changes in soil moisture content between fall, spring snowmelt, and seeding.

2. *To 'typify' the stable isotopic abundances and fluctuations of Edmonton meteoric waters.* Proper use of any tracer requires knowledge of the tracer's characteristics. As the tracers used in this study are naturally occurring, any interpretation of their abundances in the soil water and groundwater must consider their meteoric abundances and seasonal fluctuations.

3. *To determine the relative composition of soil waters and groundwaters in terms of snow and rain.* Although soil moisture measurements can be used to quantify snowmelt water entering the soil, neither mixing or displacement of the already resident waters nor quantification of the actual contribution of snow waters to the groundwaters can be evaluated in this manner. This paper uses both soil moisture measurements and isotope abundances within the soil waters and groundwaters to help achieve this objective.

4. *To determine the degree of interaction and the seasonal flow rates between the soil and groundwater systems.* This paper, divided between two chapters (Chapters 5.0 and 6.0 of this dissertation), explains the concentration of isotopes in the soil waters and groundwaters via numerical simulation. In Chapter 5.0 a one-dimensional oxygen-18 mixing model is developed and evaluated with a sensitivity analysis. The sensitivity analysis compares

changes in the output parameter, solute concentration, as the input parameters are varied. In Chapter 6.0 the model is used in combination with field data to quantify flow rates between the root zone and the water table that results in the observed isotope concentrations with depth. This sixth chapter helps to establish the link between the soil and groundwater systems.

The process of research - from that of hypothesis formulation through literature review, data gathering to interpretation of results - brings the researcher from the general to the specific. It is not until the researcher returns back to the general that a perspective can be achieved. The last chapter provides this perspective and presents a conceptual hydrological model relating the seasonal interactions of the unsaturated zone with infiltrating waters and groundwaters. This model is offered not as a final explanation for water flow phenomena, but as a basis from which future questions and investigations may begin.

1.2 STABLE ISOTOPES DEUTERIUM AND OXYGEN-18

As the chapters in this dissertation are prepared in the 'paper-format', background information for some of the techniques used has been kept to a minimum. To facilitate understanding of the use of the deuterium and oxygen-18 isotopes as tracers in the hydrological system, a brief explanation is given here.

Isotopes are atoms of a given element, but with different numbers of neutrons. Oxygen has three stable isotopes whose abundances in atmospheric oxygen are: ^{16}O=99.759%, ^{17}O=0.0374%, and ^{18}O=0.2039% (Gat 1981). Oxygen-18, ^{18}O, the isotope used for hydrological studies, has a relative abundance to oxygen-16 of 1:489 for atmospheric oxygen (Gat 1981). Ocean waters have about the same ratio 1:490 (Gonfiantini 1978). Hydrogen has two stable isotopes: ^{1}H (H) and ^{2}H (deuterium or D) with an approximate ratio of 1 D for every 6420 H atoms for ocean waters (Gonfiantini 1978). The isotopes ^{18}O and D are in sufficient abundances to be accurately measured.

As there are three stable oxygen isotopes and two hydrogen isotopes, nine different stable isotopic configurations exist for water. Of interest to us are the most common; $H_2{}^{16}O$ (18), $HD^{16}O$ (19), and $H_2{}^{18}O$ (20) with their approximate mass numbers given in brackets. As vapour pressure is inversely proportional to mass, fractionation among the different isotopes of water occurs whenever water undergoes a phase change. For evaporation, the remaining liquid becomes enriched with heavy isotopes. For condensation, rain has a greater concentration of heavy isotopes compared to the remaining vapor. Slight differences in mass between the $HD^{16}O$ and $H_2{}^{18}O$ isotopes result in different evaporation rates for these isotopes, thus making it possible to differentiate water that has undergone evaporation.

Differences in isotopic abundances can be measured far more precisely than can absolute ratios, consequently, isotope values of oxygen-18 and deuterium are reported relative to a standard. The international standard used is SMOW (Standard Mean Ocean Water). SMOW was originally a hypothetical water sample with hydrogen and oxygen isotope ratios similar to those of an average sample of ocean water (Craig 1961b; O'Neil 1986). The original waters, NBS-1A, used for defining this standard have long since been used up and the new standard, of identically similar waters, now comes from the IAEA in Vienna and is thus referred to as V-SMOW. In the literature the relative difference between the isotope sample ratio and the standard isotope ratio (δ value) is given. The δ value is defined as follows:

$$\delta_x = \left(\frac{R_x - R_{std}}{R_{std}}\right) 1000 \qquad [1.1]$$

where $R_x = (D/H)_x$ or $(^{18}O/^{16}O)_x$ of the sample being analyzed, x, and R_{std} is the corresponding ratio in the standard. R is an atomic ratio and by convention is always written as the ratio of the heavy (rare) isotope to the light (common) isotope (O'Neil 1986). The δ value is expressed in parts per thousand, or per mil (‰). A sample thus may have negative or positive δ values relative to the standard. Specification of δ values of isotopic data commonly uses $\delta^{18}O$ for oxygen-18 and δD for deuterium (O'Neil 1986).

Measurement of the isotopes, is conducted in a gaseous form, $C^{18}O^{16}O$, for oxygen and DH for hydrogen, is conducted in a high vacuum, mass spectrometer capable of measuring gases from the sample and the standard. A mass spectrometer can provide accuracy to ±0.05 ‰ for $\delta^{18}O$ and ±0.5 ‰ for δD for similar samples analyzed the same day (Gonfiantini 1981).

The $\delta^{18}O$ and δD values for precipitation, lakes, and rivers in the world range from approximately -40‰ to +10‰ for $\delta^{18}O$ and -320‰ to +40‰ for δD (Craig 1961a). For global meteoric waters, that have not undergone evaporation, a linear relationship between δD and $\delta^{18}O$, known as the meteoric water line, exists (Fig. 1.1) (Craig 1961a):

$$\delta D = 8\,\delta^{18}O + 10 \qquad [1.2]$$

This relationship has a high degree of correlation ($r = 0.997$) with a standard error of estimate for δD of ±3.3 ‰ (Yurtsever and Gat 1981). Waters that have undergone evaporation have lower δD-$\delta^{18}O$ slopes, generally between 4 and 6 instead of 8 (Craig et al. 1963) and plot to the right of the meteoric water-line (Fig. 1.1). Subsequent studies since 1961 by the IAEA, conducted at 93 stations around the world, have indicated certain trends for the δD and $\delta^{18}O$ relationship (Dansgaard 1964; Yurtsever and Gat 1981): stations at high latitudes or altitudes have more negative $\delta^{18}O$ values than do those at lower latitudes and altitudes; continental stations also have more negative $\delta^{18}O$ values than do marine stations; and for any specific station there exists a relationship between air temperature and isotopic abundances in the meteoric waters. The depletion of isotopes in rain with altitude and with increasing latitude is a function of decreasing air temperature and 'rain-out' of the heavy isotopes as the air masses move inland from the ocean.

For continental North America, the $\delta^{18}O$-δD relationship for long-term weighted mean values is (Yurtsever and Gat 1981):

$$\delta D = (7.95 \pm 0.22)\,\delta^{18}O + (6.03 \pm 3.08) \qquad [1.3]$$

for the stations; Bethel, Waco, Flagstaff, Chicago, Goose Bay, Edmonton, Fort Smith, and Whitehorse.

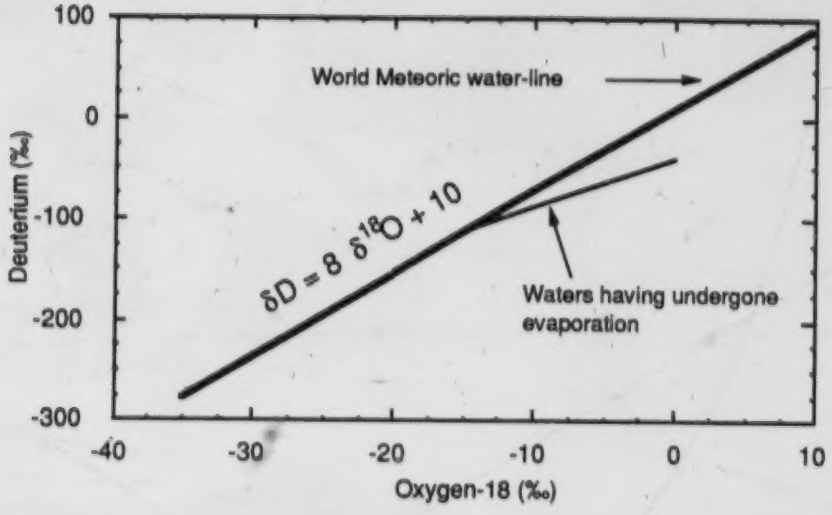

Fig. 1.1 World meteoric water-line (Craig 1961a). (adapted)

The relationship between isotope concentration and monthly air temperature is linear:

$$\delta^{18}O = (0.521 \pm 0.014)T - (14.959 \pm 0.208) \qquad [1.4]$$

where T is in degrees Celsius, for the stations Thule, Groennedal, Nord, and Vienna with a correlation coefficient of r = 0.893 (Yurtsever and Gat 1981). For continental stations the seasonal temperatures are more pronounced and thus there is a greater difference in isotopic concentrations between summer and winter seasons.

Due to the strong linear relationship between $\delta^{18}O$ and δD in meteoric waters, only one of the isotopes needs to be measured, however, for soil and lake waters where evaporation has occurred both need to be measured. During the evaporation process the lighter isotopes are more likely to be evaporated than the heavier isotopes. This results in the waters remaining becoming enriched relative to the heavier isotopes. Due to the slight mass differences between the HDO and the $H_2{}^{18}O$ isotopes there is also

slight differences between the evaporation rates of these molecules - it is for this reason that there is a shift in the slope and intercept of the δD-$\delta^{18}O$ line. If only one of the stable isotopes were measured then one could not discern whether the value of the isotope has been enriched due to evaporation. When both are measured, waters that have under gone evaporation plot to the right of the meteoric water line. For free water surfaces; air temperature, relative humidity, the isotopic composition of atmospheric water vapour, and the kinetic fractionation effect (the difference between rates at which the different isotope species diffuse into the atmosphere from an evaporating surface) affect the isotopic enrichment process (Gat and Tzur 1967; Gat and Levy 1978; Allison and Leaney 1982). Evaporation of waters from soil is complicated by soil transportation processes (Allison 1982). This can result in the slope of the $\delta^{18}O$ - δD becoming as low as 2 to 3 (Dincer et al. 1974; Allison 1982). These low slopes only occur for very dry conditions and generally are limited to the surface 10 to 20 cm of the soil profile where vapour movement by evaporation is dominant. Soil waters enriched in heavy isotopes by evaporation, may diffuse downwards even against an upward flow (Zimmermann et al. 1967). Fractionation of isotopes in soil waters is not affected by transpiration as movement to the roots is in liquid form (Zimmermann et al. 1967).

The information of isotopes in this chapter is meant to provide a general background. In the following chapters more specific information for the use of isotopes for discerning the composition of waters and the effects of evaporation are discussed.

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2. THE EFFECTS OF TILLAGE UPON SNOW COVER AND SPRING SOIL WATER.

2.1 INTRODUCTION

Approximately 30% of the annual precipitation on the Canadian prairies occurs as snowfall during the winter. Conservation and management of winter precipitation has the potential to improve crop yields. Snowtrapping techniques such as standing stubble, alternate height windrows, deflector strips, etc. have proven effective in retaining more snow than fallow fields (Lal and Steppuhn 1980; Steppuhn 1980; Nicholaichuk et al. 1984). However, the water from the augmented snow does not always enter the soil root zone, contributing instead to surface runoff (Willis et al. 1969; Chanasyk 1986). The low infiltration capacity of frozen soils imposes definite limits to water intake in many locations, especially on medium-textured soils (McConkey et al.1988).

The two controlling factors of infiltration rates in frozen soils are the frozen water content and the size and continuity of the conducting pores. Frozen soil water blocks pores and increases tortuosity. An inverse relationship between the frozen water content and infiltration rates exists with ice blockage reducing infiltration rates by 10 to 300 times (Hobbs and Krogman 1971; Murray and Gillies 1971; Kane and Stein 1983; Granger et al. 1984).

Fall cultivation of agricultural fields is practiced on the prairies in order to incorporate crop residues into the soil, to control weeds, to disperse fertilizers and herbicides, to prepare for the land spring seeding, and to promote the infiltration and soil storage of precipitation (Lal and Steppuhn 1980). Fall discing generally reduced

overwinter storage relative to that of standing stubble (Staple et al. 1960). A 9-year study in southern Saskatchewan concluded that there was no significant difference in overwinter water storage between fall blade tillage and standing stubble (Staple et al. 1960). Kachanoski et al. (1985), summarizing 4 years of measurements on different slope positions, found that fall cultivation resulted in less recharge than did standing stubble. An extensive literature review of fall tillage on the Canadian Prairies by Lal and Steppuhn (1980) indicated that tillage does not seem to increase soil water storage of overwinter precipitation as compared to stubble. Only weak trends were present with one-way disc tillage conserving slightly less water and blading slightly more than no tillage at all, especially during a heavy snow year. Chanasyk (1986) reported variable water gains from cultivated stubble as compared to standing stubble. Fall chiselling or spiking has been reported to contribute more overwinter recharge than conventional stubble (Masse and Siddoway 1969) or to have a higher recharge efficiency (Zuzel and Pikul 1987) but with less trapped snow (Patterson 1984). Subsoiling has also been reported to increase snowmelt recharge efficiency in central and southern Saskatchewan (Patterson 1984; McConkey et al. 1988).

Although some of the above studies have shown that snowmelt recharge into frozen soils may be enhanced by tillage, trends are not consistent and are likely influenced by the snowtrapping ability of the remaining stubble and the meteorological conditions of the study area during winter. This study compares the effects of four commonly used agricultural management practices on the prairies: standing stubble, disced stubble, chiselled stubble, and summerfallow upon the amount of snow accumulated and the amount of snowmelt recharge at one location in the Black soil zone of Alberta.

2.2 MATERIALS AND METHODS

2.2.1 Site description

The study site was located approximately 15 km south of the Edmonton city center at the Ellerslie Research Station. A fully equipped meteorological station is situated at the site. The study period extended from fall 1985 to spring 1987.

The climate of the Edmonton area is subhumid and is characterized by relatively warm summers and cold winters. The average annual temperature is 3°C, with January the coldest month at -14°C and July the warmest month at 16°C. The mean annual precipitation is 460 mm with 60% occurring during May through August, with July having the greatest amount (85 mm). The mean annual overwinter precipitation (November 1 through March 31) is 89 mm, or about 19% of the annual precipitation.

A large portion of the study region is under mixed grain cultivation, with barley being the common grain crop. The local topography is gently rolling to rolling. The study site is located at the toe to mid-slope of a 1-2% west slope. Parent material consists of fine-textured glacial lacustrine over morainal material. Soils in the well-drained to moderately well-drained positions of the Ellerslie Research Station, which include the study site, have been described as Eluviated Black Chernozems. Soil textures across the study area range from clay through a clay loam to a silt loam. Total carbon of the Ap horizon has been measured at 5.2% (Crown and Greenlee 1978).

2.2.2 Experimental design

To account for soil variability and site characteristics that could affect snow accumulation or infiltration, a randomized complete block design was used. This

consisted of four treatments (standing stubble, chiselled stubble, disced stubble, and summerfallow), replicated three times, resulting in a total of 12 plots. Each plot was 23 m x 45 m and included a 5 m buffer border, resulting in an effective plot size for sampling of 13 m x 35 m. The entire site had been in barley for at least three years prior to plot establishment in the fall of 1985. All plots, except the summerfallow ones, were cultivated in mid-October 1985. The summerfallow plots were left as standing stubble for the first winter, 1985-86 and fallow was accomplished by discing during the summer and fall of 1986. Thus the fallow treatment could not be properly incorporated into the statistical design until the fall of 1986, the second winter.

Chiselling was accomplished with a cultivator equipped with 4-5 cm wide shanks spaced approximately 20 cm apart and set to a 15 cm depth. The disc was a tandem double disc one with 25 cm blades spaced 20 cm apart. The discs disturbed the top 8 cm of soil.

Seed bed preparation took place during early May and consisted of shallow discing. The plots were seeded during mid-May to the end of May. Spraying for Canada thistle was done with 2-4D (Gleen by Hoegrass) at midsummer. The summerfallow plots were disced three times during the growing season to control weeds. Harvest took place during the first half of September, 1986 and 1987 and the plots were cultivated with the same management practices as the previous fall.

Five aluminum neutron tubes were installed at randomly chosen locations within each plot. To allow for proper field maintenance, they were installed after fall cultivation (mid to late October) and removed before seeding (mid to late May). New sites were

randomly chosen for each installation. Soil water was measured at 10-cm depth intervals from 15 to 95 cm on a monthly basis during winter and more frequently during melt periods using a Campbell Pacific Nuclear 503 Depthprobe.

Snow stakes, marked in two cm increments were driven into the ground 1.5 m from each neutron access tube. Readings of snow depth were obtained at the time of each soil water measurement during the winter. Snow cores for the determination of snow water equivalence (SWE) were taken four or five times each winter with a 15 cm diameter PVC tube. Three to five cores were taken from each plot.

Soil samples, from cores obtained during the first neutron access tube installation (fall 1985), were taken at 15 cm intervals from the surface to a depth 90 cm. The soil samples were analysed for bulk density, particle size using the hydrometer method (McKeague 1978), and water retention by pressure plate apparatus (Richards 1965).

2.2.3 Hypotheses tested

The following null hypotheses were tested with analysis of variance for a randomized block design with subsampling:

1. The type of fall cultivation treatment does not have an effect upon snow depth accumulation.
2. Fall cultivation does not have an effect upon the amount of snowmelt recharge.
3. Fall cultivation does not have an effect upon soil water content at time of seeding.

If the null hypothesis was rejected at the 0.05 significance level (with the F-test), then a multiple means test (least significant difference), was used to test for differences among treatments.

2.3 RESULTS AND DISCUSSION

2.3.1 Soil physical properties

Bulk density from all 60 locations ranged from 0.92 (±0.14) Mg m^{-3} (± one standard deviation) for the uppermost 10 cm to 1.56 (±0.11) Mg m^{-3} for the 0.75-0.90 m depth interval. Soil texture varied from clay to clay loam throughout the sampled profile (0 to 100+ cm) with clay content increasing with depth. Numerous discontinuous sand lenses, 1-4 cm thick, were observed below 60 cm. The Ah horizon had an average depth of 31 (±7) cm. Water content at field capacity and permanent wilting point for the top 15 cm were 0.38 m^3 m^{-3} and 0.22 m^3 m^{-3} respectively, whereas below 15 cm, field capacity ranged from 0.40 to 0.43 m^3 m^{-3} and permanent wilting point was approximately 0.27 m^3 m^{-3}.

2.3.2 Snow accumulation

After harvest stubble heights were between 10 and 15 cm. Chiselling left a rough surface with clods 5 to 15 cm in diameter. Much of the stubble after chiselling remained standing, although at a variety of angles, and combined with the rough surface frequently resulted in the stubble tops being 20 cm higher than bottom of the clods. Discing resulted in a smoother surface and a smaller clod size, less than 5 cm, and with more than 50% of the straw being incorporated. The remaining stubble usually was left standing less than 5 cm above the soil surface.

During the first fall (September and October 1985) and winter (1 November 1985 to 31 March 1986) of the study, temperature and precipitation were near normal (Table 2.1); however, the fall of 1986 had about twice the long term precipitation (normal being 58 mm), while the winter of 1986-87 was warmer and drier than normal. Total precipitation during the first and second winters of the study were 81 mm and 50 mm, respectively.

Table 2.1. Climatic summary of study period, September 1985 to May 1987.

	1985-86		1986-87		long term mean	
Season	T (°C)	P (mm)	T (°C)	P (mm)	T (°C)	P (mm)
Sept. - Oct.	5.4	65.4	7.2	109.4	7.4	57.5
Nov. - Mar.	-8.6	80.8	-6.6	49.8	-9.9	89.0
Apr. - May	7.5	84.7	9.5	80.5	6.7	64.9
Jun. - Aug.	14.9	222.5	/	/	15.0	228.9

Major midwinter melts, in which there was substantial reductions in snow depth and in SWE, occurred twice during the first winter, 19 to 25 December 1985 and 28 February to 4 March 1986 but not during the second winter. Minor midwinter melts occurred during the second winter resulting in a gradual snowpack ablation and the development of an ice layer at the bottom of the snowpack. By the beginning of March for both winters there was an ice layer on the soil surface underlying the snowpack of the stubble, chiselled, and disced plots. Generally the fallow plots had a shallow depth of snow which was periodically removed due by melts or winds. During the first winter, snow from the disced sites had completely ablated several days before the snow from the stubble and chiselled plots had ablated (27 March 1986). During the second winter, snowmelt had begun by 23 March 1987 and by 30 March only snow on the chiselled and stubble plots and ice in tillage ruts of the disced plots remained. Snow

was completely melted by 3 April 1987. Due to midwinter melts, drifting snow, and ice layers, it was difficult to quantify the total amount of snow trapped by each plot.

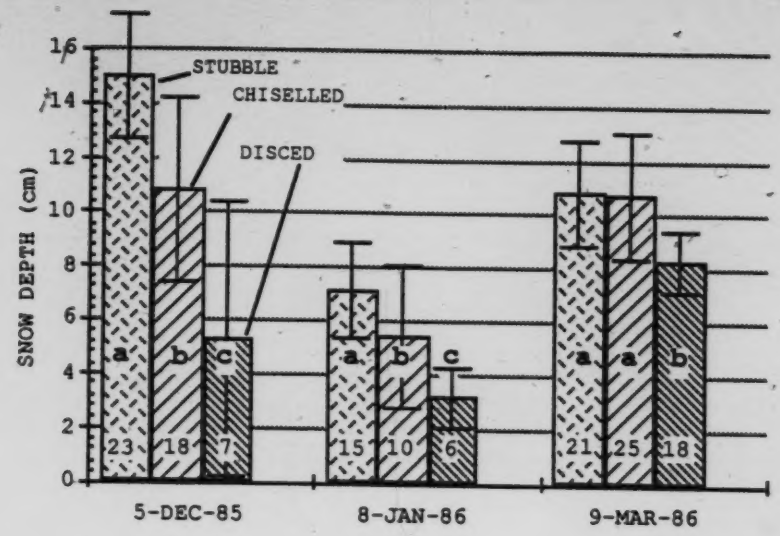
Snow depth and snow-water equivalent (SWE) for both winters are depicted in Fig. 2.1. The snow depths are the average of 15 measurements per treatment, while the SWE was measured at 3 to 5 locations per plot several times each winter. Snowfall during the first winter was greater, however, midwinter melts were more extensive and the snow amounts shown in Figure 2.1 are more representative of the snow falls between melt periods. A gradual reduction of the snow depth and SWE also occurred during the second winter; thus, the final snowpack measurements do not represent the cumulative snowfall. For both winters the greatest depth of snow was present on the stubble and chiselled plots with the smallest depth on the fallow plots. In general snow depth for the treatments may be ordered from greatest to least as follows:

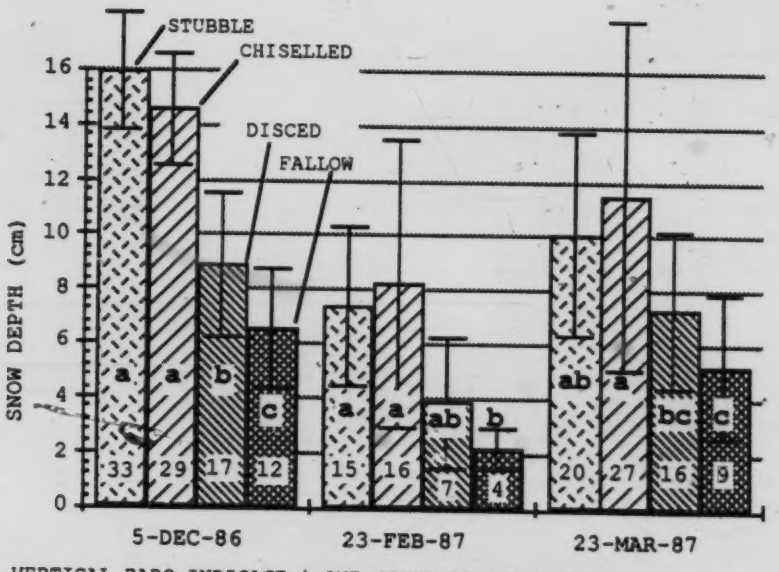
$$[\text{stubble} \approx \text{chiselled}] > [\text{disced} \geq \text{fallow}]$$

with the stubble and chiselled treatments significantly trapping more snow than the disced and fallow treatments.

2.3.3 Soil water: fall to post melt.

Fall soil water in 1985 was drier for all treatments than fall soil water in 1986, which was at or above field capacity (Fig. 2.2). A greater overwinter increase in soil water occurred during the first winter of study (1985-86) than during the second winter (1986-87) possibly due to a higher infiltration capacity due to a lower frozen water content during the first winter; a greater snowfall during the first winter; and two major melt periods occurring during the first winter. As the reduced snowfall during the second winter did not result in a proportionately lesser amount of snow





VERTICAL BARS INDICATE ± ONE STANDARD DEVIATION.
a-c: SAME LETTERS FOR SAME DATE, NO SIGNIFICANT DIFFERENCE AT 95% LEVEL, ACCORDING TO LEAST SIGNIFICANT DIFFERENCES TEST.
NUMBERS AT BOTTOM OF COLUMNS INDICATE SNOW WATER EQUIVALENT IN mm.

Fig. 2.1. Average depth and water equivalent of snow by plot treatments for the two winters of study.

measured on the plots (Fig. 2.1), and as much of the soil water increase had occurred by midwinter for both winters (Fig. 2.2, Table 2.2), the major reason for the greater increase in overwinter soil water of the first winter may have been the lower soil water contents and thus the greater infiltration capacity of the soil (Hobbs and Krogman 1971; Murray and Gillies 1971). Insufficient snow depth and SWE measurements were taken to conduct any form of water budgeting for the snowpack or the soil water.

Table 2.2. Overwinter and spring changes in soil water storage (mm) for winters 1985-86 and 1986-87.

Depth	Stubble			Chiselled			Disced			Fallow		
	Change in soil moisture storage (mm) from Nov 4 1985 to:											
(m)	27/2	31/3	29/5	27/2	31/3	29/5	27/2	31/3	29/5			
0 - 0.2	13.5	12.0	7.1	20.5	18.5	12.2	16.6	8.4	6.9			
0.2 - 0.4	3.8	16.0	21.2	6.0	19.0	20.9	4.9	13.9	14.5	not		
0.4 - 0.6	3.3	4.8	18.1	3.4	7.1	12.1	2.9	6.0	7.7	available		
0.6 - 0.8	3.9	2.4	4.7	4.0	3.9	5.4	3.7	3.6	5.2			
0.8 - 1.0	9.9	8.3	8.3	6.8	6.2	4.0	8.5	7.7	5.4			
Total	34.5	43.5	59.4	40.7	54.7	54.6	36.6	39.6	39.6			
	Change in soil moisture storage (mm) from Dec 4 1986 to:											
	23/2	6/4	9/5	23/2	6/4	9/5	23/2	6/4	9/5	23/2	6/4	9/5
0 - 0.2	6.3	7.3	-12.6	5.7	13.3	-14.8	4.7	4.9	-12.8	2.8	4.4	-13.6
0.2 - 0.4	4.7	7.0	3.3	5.7	8.0	0.5	1.2	1.5	-6.5	-1.2	0.0	-6.8
0.4 - 0.6	0.2	1.0	1.5	3.3	4.4	3.4	2.0	2.6	2.3	1.3	2.5	-0.6
0.6 - 0.8	-0.6	-0.4	3.5	0.0	1.0	3.8	1.1	2.3	2.5	2.1	2.6	1.8
0.8 - 1.0	1.1	2.9	2.8	1.9	2.3	0.8	5.3	8.7	4.6	4.4	6.6	3.6
Total	11.6	17.8	-1.5	16.7	29.1	-6.3	14.3	20.2	-9.8	9.4	16.0	-15.6

All storage numbers represent the average of measurements from 15 sites.

The only significant difference in soil water in fall 1985, among treatments, occurred at 15 cm where the water content of the stubble was greater than that of the chiselled or disced treatments. This was likely due to water loss through evaporation from the

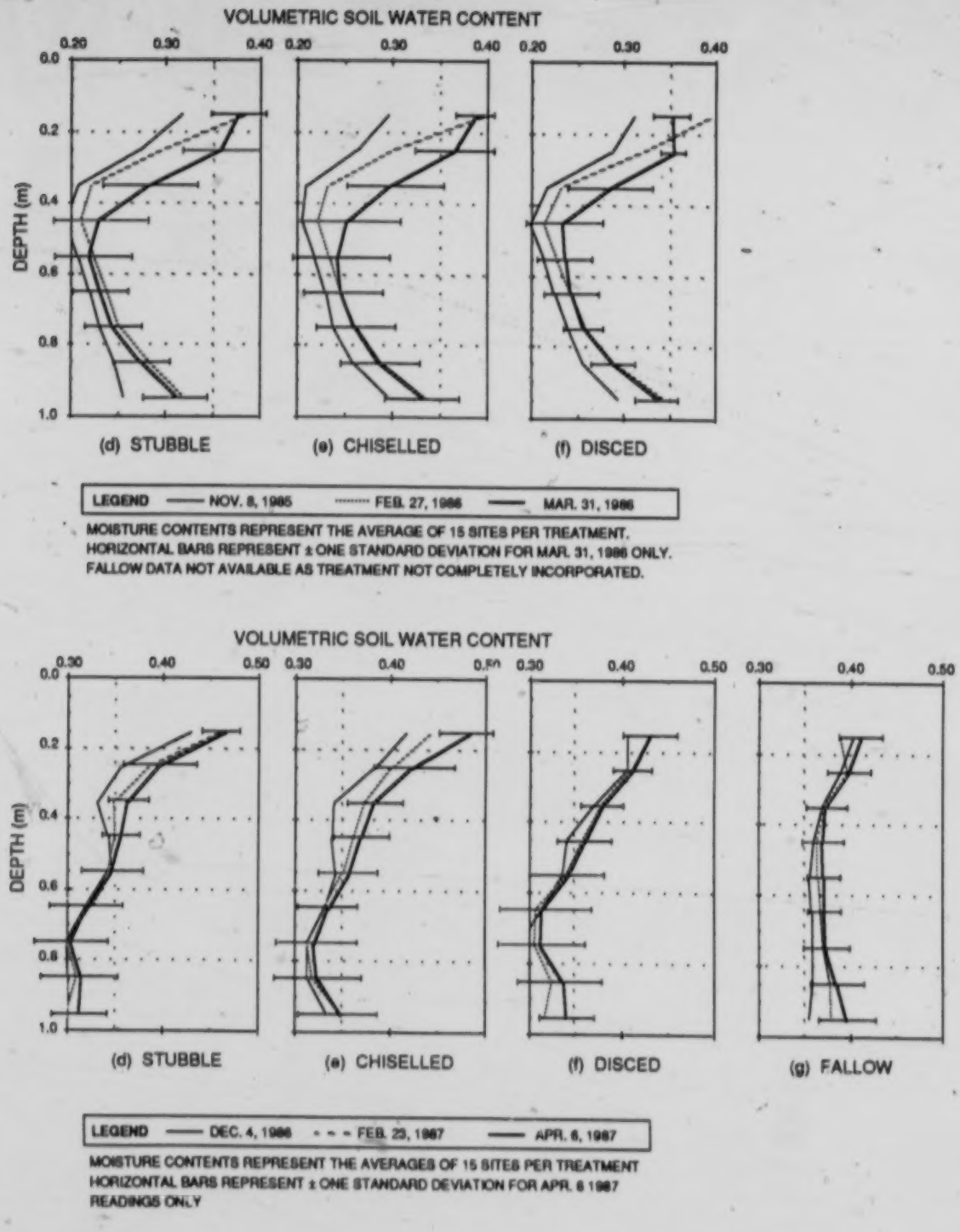


Figure 2.2. Fall to spring soil water content for tillage treatments during two winters; 1985-86 (a to c) and 1986-87 (d to g)

tilled surfaces of the chiselled and disced treatments. On 4 December 1986 the water contents of the stubble and chiselled treatments were significantly greater than those of fallow at 15 cm. This could be due to a false neutron backscatter reading at this depth due to snow on the stubble and chiselled plots. However, readings at 20 cm in November 1986 indicated that the water contents at this depth under fallow were significantly greater than those in the stubble and chiselled treatments at this depth. At 35 cm and deeper, the fallow water content during fall 1986 was significantly greater than that in the other treatments.

For the first winter, which had relatively low fall soil water contents, the chiselled treatment had a significantly greater overwinter water increase than did the stubble treatment for the 15 cm depth and the disced treatment for the 15 to 35 cm depths (Table 2.3). For the second winter, which had a wetter fall, overwinter water gains in the chiselled and stubble treatments were similar for the 15 to 35 cm depths and significantly greater than the disced and fallow treatments (Tables 2.2 and 2.3).

Table 2.3. Significant differences among treatments for overwinter soil water gains.

Depth (m)	Nov 4/85 - Mar 31/86				Dec 4/86 - Apr 6/87			
	S	C	D	F	S	C	D	F
0.15	b	a	c	/	ab	a	b	b
0.25	ab	a	b	/	a	a	b	b
0.35	ab	a	b	/	a	a	b	b
0.45	a	a	a	/	c	a	b	c
0.55	a	a	a	/	b	a	ab	a
0.65	a	a	a	/	c	b	b	a
0.75	a	a	a	/	c	c	a	b
0.85	a	a	a	/	c	c	a	b

Different letters (lowercase) designate significantly different water gains for similar depths at $P \leq 0.05$, where 'a' indicates greatest water gain, 'b' second greatest, etc.
'/' data not available.
Treatments: S = stubble, C = chiselled stubble, D = disced stubble, F = summerfallow

The soil water content immediately following springmelt was tested for significant differences ($p \leq 0.05$) among treatments. The only significant difference occurred at the 15 cm depth for both winters where chiselled and stubble treatments had significantly greater water contents than did disced (for both winters) and fallow (only the second winter could be appropriately tested) treatments.

About half to two-thirds of the fall to springmelt soil water increase had occurred by the end of February (Table 2.2 and Fig. 2.2), yet much of the snow still remained (Fig. 2.1). Much of this midwinter increase occurred in the top 20 to 40 cm. The large proportion of soil water increase that occurred by mid-winter indicates the importance of midwinter melt, but can also result in reduction of infiltration during springmelt due to increased blockage of pores by frozen soil water.

2.3.4 Soil water: post-melt to seeding.

During the period between spring snowmelt and seeding redistribution of the soil water in the profile and water gains through precipitation events, or losses due to evaporation may occur. Large changes in soil water content occurred for some of the treatments in the period between melt and seeding (Table 2.2 and Fig. 2.3). Between 31 March and 29 May 1986, there was a slight net loss from the 0 to 20 cm depth interval for the stubble treatment but below this depth there was a large gain. Between 6 April and 9 May 1987, there was a decrease in soil water from the surface to about 50 cm for all treatments (Fig. 2.3), likely due to evaporative losses. April 1987 was drier (11 mm of precipitation compared to 45 mm) and warmer (7.0 °C as compared to 3.6 °C) than April 1986.

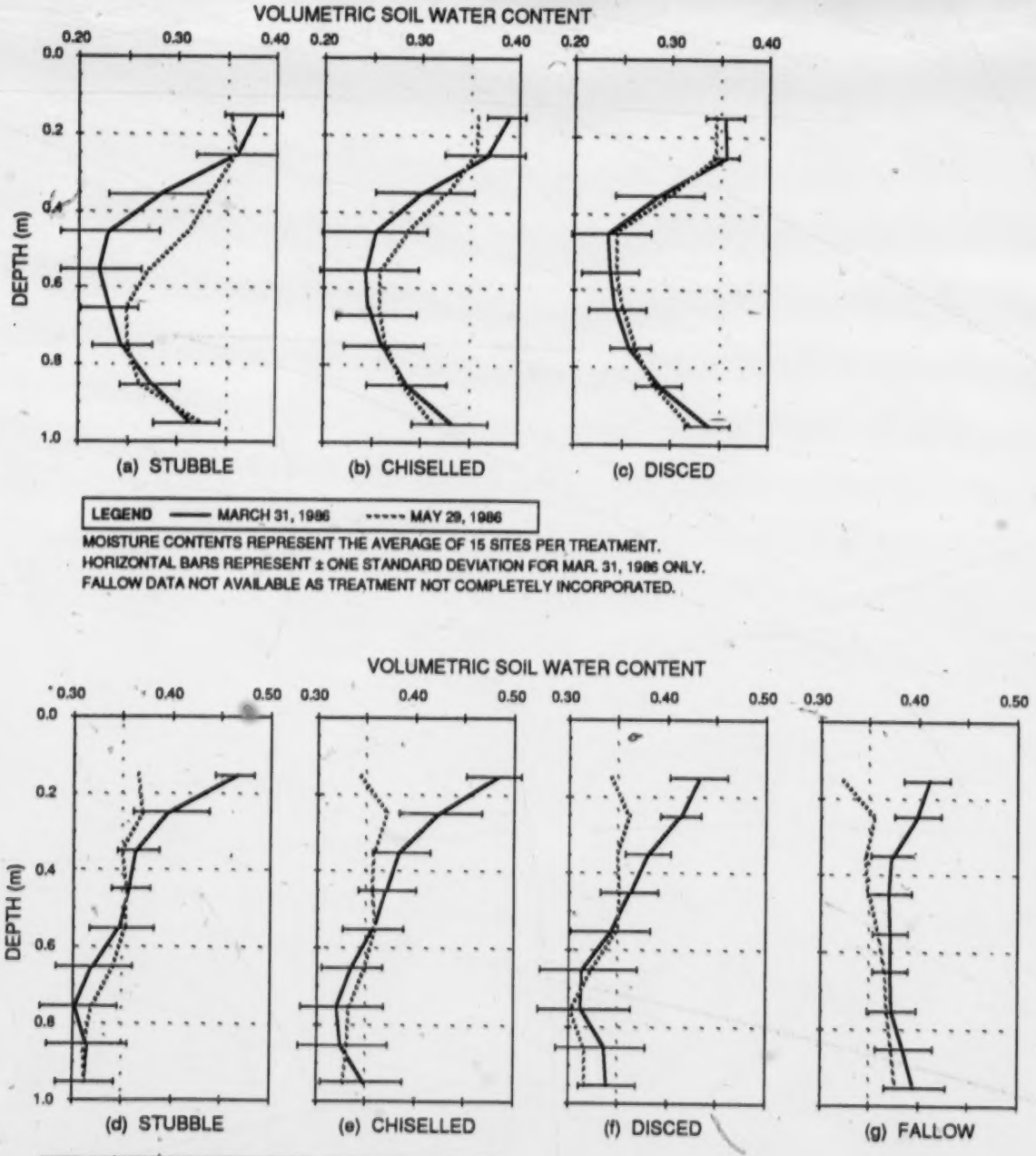


Figure 2.3. Soil water content for tillage treatments at snowmelt and seeding for two springs, 1986 (a to c) and 1987 (d to g)

The amount of water gain or loss from fall to seeding (Table 2.2) may be ordered as follows from the greatest gain (or least loss) to the smallest gain (or greatest loss):

Stubble > chiselled > disced	for the winter of 1985-86
Stubble > chiselled > disced > fallow	for the winter of 1986-87

The stubble and chiselled treatments either gained the greatest amount of soil water since fall (May 29, 1986) or lost the least (May 9, 1987) with the chiselled treatment being slightly better at soil water conservation than was stubble (Table 2.2). The 1986 stubble water contents in May were significantly greater than those in other treatments at 15 cm (Table 2.4), although the chiselled soil water contents were similar to those of stubble and significantly greater than those of disced. Soil water in fallow in May 1987 was significantly greater than soil water in the other treatments only from 65 cm and deeper (Table 2.4). The similar soil water contents in all treatments at shallower depths could be in part related to high soil water contents for all treatments during the previous fall.

Table 2.4. Before seeding soil water contents: significant differences.

Depth (m)	May 29 1986 S	C	D	F	May 9 1987 S	C	D	F
0.15	a	a	a	/	a	b	b	c
0.25	a	a	a	/	a	a	ab	b
0.35	a	a	b	/	ab	a	ab	b
0.45	a	b	c	/	a	a	a	a
0.55	a	ab	b	/	a	a	a	a
0.65	a	a	a	/	bc	b	c	a
0.75	a	a	a	/	bc	b	b	a
0.85	b	a	a	/	b	bc	b	a

Different letters (lowercase) designate significantly different moisture gains for similar depths at $P \leq 0.05$, where 'a' indicates greatest moisture gain, 'b' second greatest, etc.
'/' data not available.
Treatments: S = stubble, C = chiselled, D = disced, F = tilled summer-fallow
Fallow plots for first winter had standing stubble.

2.4 SUMMARY AND CONCLUSIONS

Four common agricultural cultivation techniques were tested to evaluate snow trapping, increases in soil water from snowmelt, and its retention until spring seeding. The techniques evaluated were three types of stubble tillage (disced, chiselled, and standing stubble) and summerfallow. The field project in two winters of study covered dry and wet fall soil conditions enabling the potential of these cultivation methods for increasing the infiltration capacity of frozen soils to be evaluated fully. The results of the study may be summarized as follows:

1. Snow depth, from greatest to least:
   [stubble = chiselled] > [disced ≥ fallow]

   Generally the stubble and chiselled treatments were not significantly different and both trapped significantly more snow than did the disced and fallow treatments ($p \leq 0.05$). Although mid-winter melts frequently

reduced the snowpack throughout the winter, the relative ranking among treatments remained the same after these melts.

2. Snowmelt recharge (fall to post-melt spring soil water gain):

0 to 0.4 m	chiselled ≥ stubble ≥ [disced ≈ fallow]	
0.4 to 0.9 m	chiselled ≈ stubble ≈ disced	for dry fall conditions
	fallow ≈ disced ≥ chiselled ≥ stubble	for wet fall conditions

For dry fall conditions, the chiselled treatment resulted in a significantly greater gain in soil water content due to snowmelt recharge at the 15 cm depth than the other treatments. For all other depths and for wet fall conditions, the chiselled and stubble treatments had similar gains. Generally the chiselled treatment always had a greater increase in soil moisture than did the disced treatment for the top 40 cm to 50 cm, whereas the stubble moisture increase was sometimes not significantly different from either the disced or the chiselled treatments. In terms of total overwinter (November-March) water gain, the chiselled treatment gained the most for both years (55 mm and 29 mm for the dry and wet years, respectively), whereas gains in the stubble (44 and 18 mm) and the disced treatments (40 mm and 20 mm) were similar.

The fallow treatment during the second winter had a slightly greater, though not significant, spring water content only below 45 cm than the other treatments. The only significant difference in spring soil water content among treatments, occurred at 15 cm where the stubble and chiselled soil water contents were greater than those for disced and fallow under wet fall conditions.

3. Water content at seeding time

0 to 0.35 m	stubble ≈ chiselled ≥ disced	for dry fall conditions
	stubble ≥ chiselled ≥ disced > fallow	for wet fall conditions
0.35 to 0.90 m	fallow ≥ [stubble ≈ chiselled ≈ disced]	

In terms of greatest water gain (or least lost) from fall to seeding, stubble gained the most (1985 to 1986) or lost the least (1986 to 1987), with chiselling ranking second. Soil water contents, however, were not significantly different among treatments, with the exception of fallow being significantly lower than stubble and chiselled for the top 30 cm to 40 cm and being significantly greater than all the other treatments below 60 cm.

Overall, soil water contents of the chiselled and stubble treatments were not different from each other except in the surface 20 cm. This difference could be due to the rough surface of the chiselled treatment which was observed to hold small pockets of meltwater and ice. On sloped conditions, less runoff would occur than in stubble or disced treatments. Also because the soil is exposed by chiselling, runoff that would occur would likely have a higher sediment load than under a stubble treatment. Also, the chiselled treatment seemed more susceptible to water loss by evaporation before spring seeding, with the end result of slightly lower to equivalent surface water contents by the time of seeding compared to the stubble treatment. The chiselled treatment or a treatment that leaves a rough surface with some stubble still standing needs to be investigated further under different climatic conditions. Perhaps the chiselled treatment is more effective than a standing stubble treatment if greater amounts of snow can be retained by the chiselled stubble or in case of high post-melt precipitation. The effect of tillage upon snowmelt recharge should be investigated more closely by quantifying the effect of tillage upon straw incorporation, clod size, depth of disturbance, and surface roughness.

The discing treatment appeared to offer no advantage in comparison to the the stubble or chiselled treatments; it resulted in significantly less snow trapped and significantly lower soil water contents at seeding. This is due to the lack of standing stubble, which might have caught snow and protected the soil from evaporation, and the smaller depth of disturbance and clod size as created by the discing. However, despite the fact that the disced treatment trapped less snow than did the stubble treatment, its overwinter moisture gain was not much less than the stubble treatment. This could be in part due to the tilled surface having a slightly better infiltration capacity. Although stubble displayed a tendency to conserve more soil water between the end of springmelt and seeding, more weed problems occurred during the ensuing growing season.

As the fallow treatment was only established for one year, it could not be evaluated for the range of climatic and soil water conditions. For the 1986-87 year of measurement, wet fall conditions resulted in the spring soil water content of the fallow treatment being similar to the other treatments for the top 60 cm.

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3. MEASUREMENT AND PREDICTION OF ENVIRONMENTAL ISOTOPES IN EDMONTON METEORIC WATERS.

3.0 INTRODUCTION

Soil water and groundwater tracers enable measurement of flow rates and directions, apparent dispersion, and residence times. The use of the environmental isotopes, deuterium and oxygen-18, have the following advantages over 'added' tracers: they occur naturally, thereby offering the potential of an historical record; they are added uniformly across the landscape with meteoric waters and not as a point source; they provide the water molecule with 'double' labels; they are non-reactive; and and they enable the study of evaporation. These isotopes have been used to differentiate groundwaters from meteoric waters (Gat 1971; Schwartz and Muehlenbachs 1979; Wallick 1981), to determine water sources in hydrological studies (Krouse et al. 1978; Saxena 1984; Lawrence 1987), and to determine infiltration and evaporation in soil water systems (Dincer et al. 1974; Barnes and Allison 1983; Bengtsson et al. 1987). Reliance upon the proper use and the correct interpretation of these environmental isotopes for soil and shallow groundwater systems requires knowledge concerning their meteoric abundances and seasonal fluctuations.

The use of isotopes in hydrologic systems is divided by Lawler (1987) into two inter-related problems: the sampling of the input signal within the meteoric waters, and the sampling of signal attenuation in the hydrologic waters. This paper is concerned with the first problem: the determination of when the input signal variation is most likely to be greatest and therefore most likely to provide a traceable signal.

There have been several hydrological and shallow groundwater studies on the prairies (Fritz and Krouse 1973; Steppuhn et al. 1975; Krouse et al. 1978; Wallick et al. 1984; Freeman 1986) although only one study (Freeman 1986) measured the yearly meteoric isotope abundances. Also, only Freeman (1986) measured both meteoric deuterium and oxygen-18.

There have been numerous meteoric water isotope studies in central Alberta. One study, part of a global inventory, was conducted from 1961 to 1966 (IAEA 1969, 1970, 1971). In this study both deuterium and oxygen-18 of monthly precipitation collections from the Edmonton Municipal Airport were measured. Brown et al. (1971), inventorying Canadian waters for possible use in heavy water fractionation, continued the deuterium analyses of monthly precipitation at Edmonton from 1966 to 1969. From 1970 to 1976, a series of studies of the isotopic abundances of individual precipitation events were conducted at the University of Alberta. Linton (1972) analyzed oxygen-18 abundances of precipitation between May 1970 and September 1971 to determine the variation of oxygen-18 during individual summer rain events. Hage et al. (1975) continued Linton's study providing additional information of oxygen-18 abundances of snow storms from December 1971 to March 1972. During the summers of 1974, 1975, and 1976 oxygen-18 data on rain, hail, and atmospheric vapour were collected as part of the Alberta Hail Project in central Alberta (Sandu 1978; Ellis 1979). Charlton (1988) measured deuterium and oxygen-18 values for vapour samples collected during August 1981 in central Alberta.

Of the above mentioned studies on meteoric precipitation, only those conducted by the IAEA and Freeman (1986) none of the above mentioned studies on meteoric precipitation analyzed both oxygen-18 and deuterium. As evaporation results in the

enrichment of the heavier isotopes in the remaining liquid it is essential that both oxygen-18 and deuterium be analyzed so as to determine the amount of evaporation. This is essential for determination of seasonal precipitation contributions to the unsaturated and shallow groundwater systems.

Whenever water condenses or evaporates, fractionation occurs between the different isotopes due to the kinetic energy of molecules which is directly affected by air temperature. Thus air temperature is significantly correlated with isotope abundances in precipitation (Yurtsever and Gat 1981). Other meteorological factors affecting meteoric isotopic abundances are the amount of precipitation (Gat 1980), the distance of raindrop fall from the cloud base to the ground (Ellis 1979; Gat 1980), and the source of the atmospheric water vapour (Hage et al. 1975; Lawler 1987).

The objective of this paper is to characterize Edmonton meteoric isotope values and fluctuations to permit their use in the interpretation of isotopic composition of soil and shallow groundwater systems. Edmonton, Alberta has the second largest variation in seasonal isotopes amongst 93 global IAEA sites (Nord, Greenland has the largest) (Yurtsever and Gat 1981). The difference in average oxygen-18 abundances ($\delta^{18}O$) between summer and winter precipitation is 12‰ (IAEA 1969, 1970, 1971). Because oxygen-18 may be measured to ± 0.05‰ (Gonfiantini 1981) this large seasonal difference makes environmental isotopes a possible tool for use as tracers in soil and groundwater studies. The objective has three aspects:

1. Measurement of both deuterium and oxygen-18 in meteoric waters. This enables one to discern alteration of isotopic concentrations in soil waters and groundwaters due to evaporation.

2. Determination of seasonal fluctuations in isotopic abundances. Knowledge of these events enables a greater accuracy in the interpretation of variations in soil and groundwater systems.
3. Development of equations to estimate meteoric isotope abundances with the use of covarying properties. As the waters comprising the soil and shallow groundwater systems may easily be up to several decades old, it is vital to be able to interpret these abundances in terms of past meteoric events.

3.2 MATERIALS AND METHODS

Seasonal designations used in this study approximate the Prairie climate as outlined by Longley (1972). Although snowmelt generally begins in the last week of March, spring is considered to be from 1 April until the end of May. Summer is considered to be from mid-June to mid-August, but due to grouping of data by months it is convenient to consider summer to be June, July, and August. Fall consists of September and October as cold temperatures and a permanent snow cover in central Alberta do not generally occur until the beginning of November. Winter is thus the months November through March.

Several data sets collected and published by other researchers have been used extensively in this paper. Many of their values were averaged to represent seasons. Additional climatic data were obtained from meteorological records and used with these data sets. Monthly isotope data presented by the IAEA (1969, 1970, 1971) and Brown et al. (1971) were from composite samples of monthly precipitation. Monthly temperatures presented within this paper were averaged only from days on which precipitation occurred; the average daily temperature was weighted according to the daily amount of precipitation. All of the IAEA's samples were at the Edmonton Municipal Airport; whereas several samples of Brown et al. (1971) were collected at the Edmonton International Airport. Daily isotope data presented by Linton (1972)

were collected in the Edmonton region. Linton (1972) measured the air temperature at the time of collection, but all other climatic information (precipitation, relative humidity) were obtained from daily records of the Edmonton Municipal Airport. Much of the daily isotope data presented by Sandu (1978) and Ellis (1979) were from meteoric waters collected while studying convective storms in central Alberta. Consequently only their climatic information, which does not include precipitation amount, is available. Some of their data were collected in the Edmonton region and for these data, climatic records from the Edmonton Municipal Airport were used. Isotope data for the winter of 1971 to 1972 from Hage et al. (1975) were obtained from Figures 5 and 10 of their paper. Air temperatures provided by Hage et al. (1975) were from radiosonde taken at the height that corresponds with the 0.080 to 0.085 MPa pressure range. Air temperatures used in this report for their isotope values were obtained from Edmonton Municipal Airport records. Seasonal averages calculated from monthly or daily isotope values were weighted according to precipitation amount. Brown et al. (1971) analyzed only deuterium. So that their values could be compared to other studies, that just analyzed oxygen-18, the deuterium values were converted to oxygen-18 by using the δD-$\delta^{18}O$ relationship established from the IAEA data.

In the current study 46 samples of rain, snow, and hail representative of individual precipitation events between January 1986 and February 1988 were collected and analyzed for oxygen-18 and deuterium abundances. Thirty-six rain samples were collected with a use of the standard rain gauge located within the city, 0.5 km south of the University of Alberta campus. Generally only rain events greater than 5 to 10 mm per day were collected. Six snow samples for individual events were obtained from a raised surface (to prevent accumulation of drifting 'old snow') cleaned of previous snowfall. Other snow samples, representative of the entire snow profile, were taken

using snow cores from Ellerslie, 15 km to the south of campus and thus cannot be dated to a specific event. Climatic information was obtained from the Edmonton Municipal Airport located 5 km to the north of campus.

Both deuterium and oxygen-18 were analyzed using facilities at the University of Alberta. For isotopic analysis of oxygen-18, 5 ml of water were equilibrated with CO_2 at 25°C for 24 hours followed by isotopic analysis of the CO_2 (Epstein and Mayeda 1953; Gonfiantini 1981). Analysis of deuterium involved reduction of 15 to 20 µL of sample in an evacuated tube with 0.5 g to 0.7 g of Analar zinc shot heated to 450 °C (Coleman et al. 1982). This produced hydrogen gas which was measured by a VG Micromass® 602C. Concentrations of oxygen-18 were determined using a VG Micromass® 602D. Three international standards, V-SMOW, GISP, and SLAP, were also analyzed to properly calibrate the samples analyzed.

3.3 RESULTS AND DISCUSSION

3.3.1 Anomalous Data

Some of the data from previous studies were not included in the statistical analysis of this section due to inconsistences. For the IAEA (1969, 1970, 1971) one value was not included; February 1965 ($\delta^{18}O$ = -43.4 ‰, δD = -228.1 ‰) was omitted as the oxygen-18 value was too low and did not fall on the meteoric water line (a $\delta^{18}O$ value of -43 ‰ would have had a δD value of -323 ‰ if it was on the meteoric water line as in Fig. 1). The deuterium data collected by Brown et al. (1971) were slightly higher for winter abundances and slightly lower for summer values compared to other data (Tables 3.2, 3.3, and 3.5, Fig. 3.2), and although presented, were not utilized for statistical or predictive purposes. Such a discrepancy with the data presented by Brown et al. (1971) could be due to incorrect calibration with standards.

3.3.2 The Meteoric 'Waterline'

To adequately use the meteoric water line for studies it is important that it be properly quantified and any deviations from expected be explained. The 1986 to 1987 Edmonton meteoric water line has a slightly lower slope and intercept than that of the world meteoric water line (Fig. 3.1) and other water lines (with the exception of Fort Smith, North West Territories) including the Edmonton water line established by IAEA during 1961 to 1966 (Table 3.1). It is typical for inland continental regions to have lower coefficients than coastal regions or the global meteoric line as the rain has been depleted from ocean values due to rain-out and because atmospheric vapour is commonly less saturated than maritime environments (Yurtsever and Gat 1981). In North America, stations of increasing altitude and latitude have decreasing values of $\delta^{18}O$ and δD for equivalent temperatures. This coincides with smaller mean annual $\delta^{18}O$ values (Yurtsever and Gat 1981).

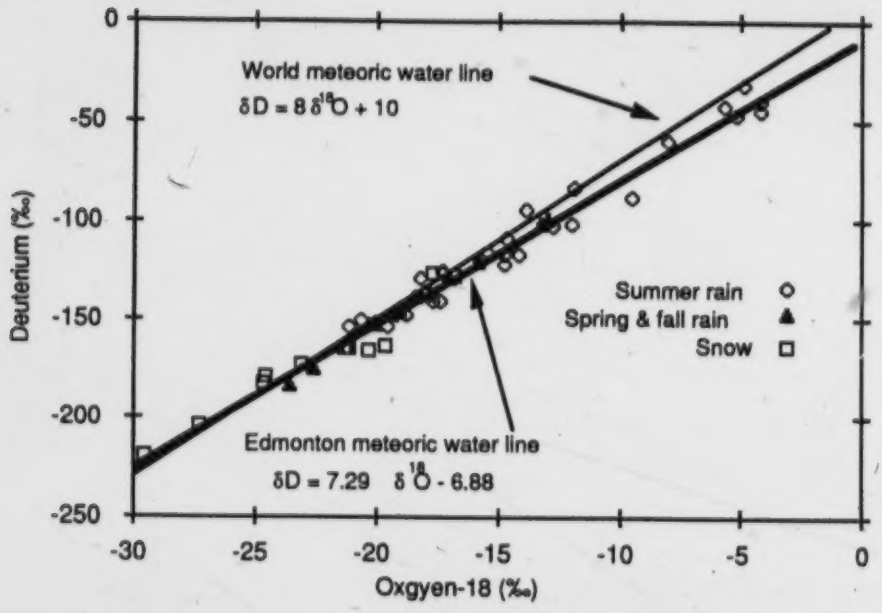

Fig. 3.1. Deuterium and oxygen-18 isotopes in Edmonton meteoric waters.

Table 3.1. Meteoric water lines

Site	slope (±SD)	intercept (±SD)	n	r^2	reference
Global	8.17 (0.08)	10.56 (0.64)	74	0.997	Yurtsever and Gat 1981
North America	7.95 (0.22)	6.03 (3.08)	8	0.997	Yurtsever and Gat 1981
Cypress Hills	7.72 (0.08)	1.98 (1.96)	24	0.999	Freeman 1986
Edmonton 61-66	7.66 (0.15)	-1.00 (3.03)	51	0.991	IAEA 1969, 1970, 1971
Fort Smith	6.8	-20.9		0.98	IAEA data (Hoyt et al. 1975)
Edmonton 86-87	7.29 (0.15)	-6.88 (2.67)	52	0.989	1986-87
... summer	7.23 (0.24)	-7.51 (3.61)	29	0.985	1986-87
... winter	6.61 (0.44)	-23.22 (9.86)	12	0.978	1986-87
... summer†	7.09 (0.25)	-9.86 (3.75)	28	0.985	1986-87
... winter*	6.91 (0.24)	-14.46 (5.47)	10	0.995	1986-87

† violent thunderstorm on July 31, 1987 removed ($\delta^{18}O$ = -4.7 ‰)

* two snowbank samples on March 23, 1987 removed ($\delta^{18}O$ = -19.7 ‰, -20.3 ‰)

Hage et al. (1975), in a re-analysis of data collected by the IAEA on Edmonton, Fort Smith, and Whitehorse, found that the slope of the oxygen-18 relationships differed appreciably between summer and winter (summer was lower) for Fort Smith and Whitehorse but not for Edmonton. Hage et al. (1975) suggested this could be due to different histories of seasonal water vapour originating from the Pacific Ocean, but this still fails to explain the negative intercepts. Almost all the warm season precipitation at these stations originates from water vapor produced by re-evaporation of precipitation over land (including transpiration), whereas at low temperatures, precipitation originates mainly from unaltered ocean vapour (Hage et al. 1975). The lower slope and intercept of the the more recent Edmonton data suggests the atmospheric water vapour supplying the precipitation was slightly more depleted than the precipitation analyzed by IAEA. This depletion, is also characteristic of precipitation events of small amounts (less than 5 mm) that have undergone evaporation during fall. As the 1986 to 1987 data did not include daily precipitation amounts less than 5 mm, whereas the IAEA data did, it is unlikely that this shift is due to the 'amount' effect. If so it the shift in the meteoric line would have been opposite to what occurred.

Summer and winter meteoric water lines for Edmonton 1986 to 1987 (Table 3.1) have similar slopes and intercepts if several samples, slightly offset from the water line, are removed. Although these samples are not distinct outliers there is reason to suspect that they might have undergone processes not typical of normal precipitation. The thunderstorm July 31 1987 was unusual for its size and intensity of rain. The two snowbank samples could have been enriched from partial melt and/or sublimation. With these points removed, the differences in equations between summer and winter could be due to sample size and measuring error as indicated by the standard deviations

of the slope and intercepts. The differences could also reflect water vapour origins and histories between summer and winter. Hage et al. (1975) stated that summer water vapor is from re-evaporation of precipitated waters over land whereas winter water vapour is from ocean vapour. The winters 1985-86, 1986-87, and 1987-88 were, respectively, 2.0, 3.6, and 4.7 °C warmer than the long term average at the Edmonton Municipal Airport (-8.5 °C). These abnormally warmer and drier winters could be the result of different weather patterns which could also affect water vapour origins and histories.

3.3.3 Seasonal variation of oxygen-18.

The average annual temperature is 3.5 °C (± 1.3 °C) with average summer temperatures being approximately 17°C (± 0.7) and average winter temperatures at -8.5 °C (±2.7). The average annual precipitation at Edmonton Municipal Airport is 457 mm (±75 mm), with slightly more than half of this (233 ± 47 mm) falling during the three summer months. The average winter precipitation is 95 (± 30.4) mm. There are usually one or two midwinter melts resulting in some snowpack reduction but most of the melt occurs during spring starting in the last half of March.

The most comprehensive isotope data set for the Edmonton region is that collected by the IAEA between 1961 and 1966 (Table 3.2). According to the IAEA, data the average winter oxygen-18 value differs from the average summer value by about 12 ‰. For the other data sets, the differences ranged from 12 to 16.5 ‰; however, with the exception of the summers 1971 and 1987, these data sets are not considered representative due to their small sample size. The isotope values for summers of 1971 (Linton 1972) and 1987 are from rainfalls that represent 67% and 91% of the total summer rainfalls respectively. Isotope values for the summer of 1986 represent 57% of the total summer rainfall. If the year is differentiated primarily on form of

precipitation, snow and rain, then the weighted oxygen-18 average (weighted by rainfall amount for those days that rain occurred) for the IAEA rain is -15.2 ‰ making the difference between snow and rain still quite large (10.7 ‰). The summers of 1986 and 1987 had extreme mean $\delta^{18}O$ values for all the collected data; -17.7 ‰ for 1986 to -13.3 ‰ for 1987.

Table 3.2. Seasonal oxygen-18 averages (weighted according to precipitation amount).

Years	Winter ‰ (sd) n	Spring ‰ (sd) n	Summer ‰ (sd) n	Autumn ‰ (sd) n
1961-1966†	-25.9 (4.0) 15	-19.2 (3.9) 8	-13.8 (1.7) 15	-15.6 (3.1) 8
1966-1969†	-22.1 (3.5) 12	-19.1 (4.0) 4	-16.7 (2.2) 9	-16.4 (1.4) 4
1970	-30.5 (0.7) 2	-18.1 1	-14.0 (2.3) 8	-11.0 1
1971‡	-29.3 (3.5) 8	-13.4 (0.3) 2	-17.0 (2.2) 16	-19.7 (0.2) 2
1974		-8.9 1	-15.2 (3.6) 9	
1975			-13.5 (5.1) 8	
1986		-16.7 (3.2) 6	-17.7 (3.4) 8	-20.8 (3.0) 4
1987	-25.3§ (5.8) 5		-13.3 (5.1) 17	
1970-87	-27.9 (4.4) 16	-16.2 (3.1) 10	-15.2 (4.1) 67	-19.0 (3.8) 7

† - isotope values represent monthly precipitation samples, remainder from daily data.
‡ - most of winter data is from winter of 1971-72, with 2 points from winter 1970-71.
§ - data from winter of 1987-88.

3.3.4 Prediction of Oxygen-18 from Climatic Parameters

1. Annual data

Due to the strong relationship between air temperature and isotopic fractionation, there also exists a strong positive correlation between air temperature and isotope abundance (Yurtsever and Gat 1981). This is confirmed by the Edmonton isotope data (Fig. 3.2, Table 3.3); however, the standard error is relatively high, between 2 and 3‰ for oxygen-18 (Table 3.3).

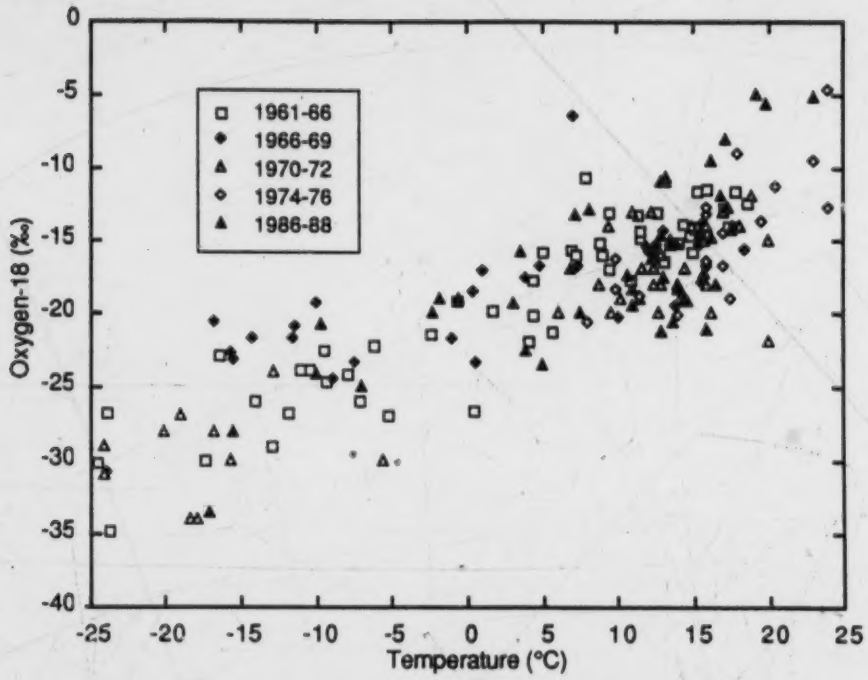

Fig. 3.2. Oxygen-18 (‰) values of Edmonton meteoric waters and air temperature.

Table 3.3. Correlation values for temperature (T) and $\delta^{18}O$

Dates	n	Regression equation intercept	slope	r^2	prob. >F	Standard error
Annual						
1961-66	50	-20.40 (0.34)	0.442 (0.03)T	0.920	0.000	2.37
1966-69	29	-19.66 (0.41)	0.251 (0.03)T	0.828	0.000	2.19
1970-72	38	-21.65 (0.50)	0.411 (0.03)T	0.895	0.000	2.78
1986-88	43	-21.14 (0.70)	0.427 (0.06)T	0.774	0.000	3.66

With the exception of the data collected by Brown et al. (1971) for 1966 to 1969, the regression equations in Table 3.3 are similar with slopes between 0.41 and 0.44 and an intercept of approximately -21 ‰. The regression equation for 1961 to 1966 has the highest slope and intercept which results in slightly higher estimates of $\delta^{18}O$. This could possibly be due to the fact that the IAEA data include low amounts of

precipitation (< 5 mm/day) that were not sampled for the daily data sets. Low rainfalls are more likely to become enriched due to evaporation while falling. Of rainfall amounts for the five summers during 1961 to February 1966, 23.7% of the rainfall was less than 5 mm, whereas for the 1970 to 1971 and 1986 to 1987 summers only 6.5% and 12.2% respectively of the rainfall events sampled were less than 5 mm. Thus the equations representing individual events are regarded as being more representative of rains likely to infiltrate into the root zone and groundwater.

Although the isotope-temperature relationship is frequently represented by a linear equation, Hage et al. (1975) using Edmonton data collected between 1970 and 1972 described the $\delta^{18}O$-temperature relationship as curvilinear in which cold and warm temperatures (< -20 °C and > 5°C respectively for a 0.085 MPa altitude) had flat slopes near zero, while mid-range temperatures had steeper slopes. A similar visual analysis for the 1970-88 data set indicates that the $\delta^{18}O$-temperature slope for the mid-range temperatures (-5 to 15 °C) is low, while the slope for the higher temperatures (15 to 25+ °C) is steeper (Fig. 3.3), opposite to that described by Hage et al. (1975). Winter data for 1970 to 1988 indicate a 'levelling' trend which is consistent with that described by Hage et al. (1975). As the standard deviations of the 'grouped' data are quite large relative to the 'curvilinear' trends shown in Fig. 3.3, caution must be taken concerning the actual interpretation of this or any data as being 'curvilinear'.

Hage et al. (1975) stated that the reduced slope at higher temperatures is due to poor correlation between the isolated convective rain storms and the air temperatures taken by the radiosonde 50 km from the rain collection site. Perhaps another reason for the difference between the high temperature data sets is the lack of $\delta^{18}O$ values greater

than -10 ‰ for the Hage et al. (1975) data. For the data collected during the summers 1974 to 1976 and 1986 to 1987, there were eight precipitation events with values higher than -10 ‰ (Fig. 3.2). These higher values seem to be associated with temperatures slightly lower than that predicted by the linear equations in Table 3.3, as noted by the steep slopes in the 15+ °C temperature range of Fig. 3.3.

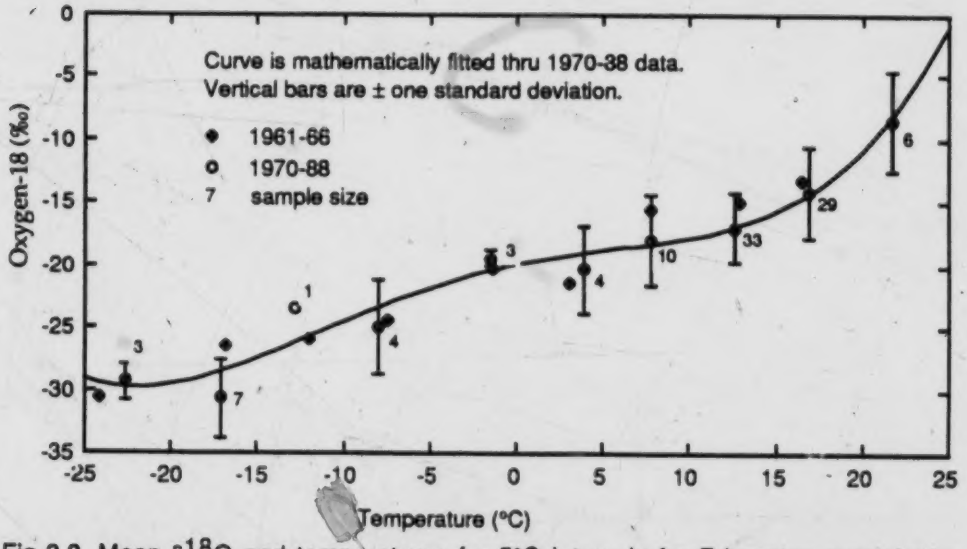

Fig 3.3. Mean $\delta^{18}O$ and temperatures for 5°C intervals for Edmonton precipitation.

The curvilinear trend described in Hage et al. (1975) and in Fig. 3.3 is not strong. This is evidenced by the large standard deviations. Also some data points may be affected by small sample sizes at some temperatures particularly at the end of the curve (Fig. 3.3). However, if the $\delta^{18}O$-temperature data are subdivided according to three temperature divisions suggested by Fig. 3.3, ≤ 0°C, -5 to 15 °C, and ≥ 15°C, then different correlation values emerge (Table 3.4). For temperatures less than 0 °C both the IAEA data and the individual event data were used as precipitation at these temperatures, being solid in form, is not affected by evaporation and exchange during fall. The values for rain (above 0°C) utilized only the daily isotope data (1970 to 1987) as it was felt that the IAEA data, which included daily rain amounts less than 5

mm/d, would have biased the results away from that representative of waters that infiltrated into the soil.

Table 3.4. Correlation values for $\delta^{18}O$ and temperature.

Temperatures Sample dates	n	Regression equation intercept	slope	r^2	prob. >F	Standard error
T ≤ 0°C						
1961-88	34	-20.55 (1.01)	0.457 (0.07)	0.747	0.000	2.96
T >-5°C, ≤15°C						
1970-88	50	-19.56 (1.05)	0.208 (0.09)	0.303	0.034	2.83
T > 15°C						
1970-88	35	-34.81 (4.26)	1.225 (0.24)	0.664	0.000	3.18

2. Summer and winter data

The seasonal $\delta^{18}O$ means in Table 3.2 and the correlation coefficients in Table 3.5 indicate large differences not only between seasons but from year to year. Some of these differences are probably due to the lack of sufficient sample size to represent the season adequately. This is more noticeable for the $\delta^{18}O$-temperature slopes of the winters, 1970-72, and 1986-88 which have extremes of 0.098 and 0.73 respectively. Summer values also vary widely with slopes ranging from 0.17 (1971) to 1.37 (1986). Variation with winter values may be due to the different origins of atmospheric water vapour, whereas variation in summer values due to evaporation and exchange processes with atmospheric vapour of the rain during its fall to the ground in addition to the different origins of water vapour.

Table 3.5. Seasonal correlation values for temperature and $\delta^{18}O$

Dates	n	Regression equation intercept	slope	r^2	prob. >F	Standard error
June - Aug						
1961-66	15	-22.17 (3.13)	0.544 (0.20)	0.596	0.019	1.38
1966-69	9	-23.67 (4.33)	0.477 (0.29)	0.525	0.138	1.96
1970	8	-22.13 (4.24)	0.527 (0.27)	0.620	0.099	1.96
1971	16	-19.28 (2.86)	0.174 (0.21)	0.213	0.428	2.20
1974	9	-22.14 (3.98)	0.437 (0.24)	0.563	0.109	3.16
1975	7	-31.33 (3.49)	0.998 (0.19)	0.920	0.003	2.34
1986	8	-38.60 (7.73)	1.375 (0.51)	0.743	0.035	2.46
1987	17	-32.15 (4.00)	1.220 (0.25)	0.779	0.000	3.29
Nov to March						
1961-66	30	-20.92 (1.19)	0.395 (0.09)	0.753	0.000	2.54
1966-69	12	-17.94 (1.34)	0.371 (0.10)	0.734	0.005	2.41
1970-72	10	-27.83 (3.72)	0.098 (0.21)	0.166	0.646	3.31
1986-88	8	-18.07 (1.79)	0.734 (0.16)	0.878	0.004	2.43

The effect of water vapour upon isotope concentrations within precipitation is more effectively studied if snow or hail data is included with rain. This is because evaporation and exchange between solid precipitation and water vapour is minimized during fall. Hage et al. (1975) found that snow could be used as an indicator of whether water vapour was of marine origin or from re-evaporated precipitation. Marine vapour is attributed to a $\delta^{18}O$ snow value of -23 ‰ for a mean daily temperature of -12.8 °C (9 March 1972), whereas a snowstorm typical of re-evaporated precipitation (15 January 1972) had more a more depleted $\delta^{18}O$ value of -30 ‰ for a higher temperature of -5.6 °C (Hage et al. 1975). An 11 ‰ range in snow $\delta^{18}O$ values during snowfall between 2 and 6 February were partly attributed to water vapour of different origins (Hage et al. 1975).

For central Alberta almost all warm season precipitation originates from water vapour produced by re-evaporation of local water (Hage et al. 1975; Sandu 1978). Occasionally maritime water vapour from the Pacific ocean that has not been

isotopically altered by previous rains may come to Alberta. Normal $\delta^{18}O$ water vapour values for the summer months are usually approximately -26 ‰ (based upon 66 samples, Ellis 1979). A high $\delta^{18}O$ value for water vapour (-9.5 ‰) is attributed to such maritime air by Ellis (1979). Actual ocean vapour, averaged from numerous samples, has a mean $\delta^{18}O$ value of -13 ‰ and rains from this vapour have $\delta^{18}O$ values of -4 ‰ (Craig and Gordon 1965; Yurtsever and Gat 1981). A storm on July 3, 1975 had hail with $\delta^{18}O$ values ranging from -13.8 ‰ to -10.8 ‰ and with rain $\delta^{18}O$ values of -5.3 ‰ and -3.9 ‰ (Sandu 1978). The vapour feeding the storm was estimated to have a $\delta^{18}O$ value of about -7 ‰ (Sandu 1978). A violent thunderstorm on July 31, 1987, resulting in large hail stones (>3 cm dia.) and a tornado, had a $\delta^{18}O$ rain value of -4.9 ‰ and a hail $\delta^{18}O$ value of -11.9 ‰ . Water vapour in this storm likely was unaltered ocean vapour.

For Albertan summer storms, evaporation of the falling raindrop can result in enrichment of $\delta^{18}O$ values by 3 to 4 ‰ for a one kilometer fall (Linton 1971; Ellis 1979). This enrichment; however, is only dominant for rainfalls of low amounts or the initial rains of a storm while the air is still unsaturated (Ellis 1979; Yurtsever and Gat 1981). With larger amounts of rainfall or continual rainfall the atmosphere becomes more saturated with water vapour from the storm and isotopic exchange ceases to have such a dominant effect. Ellis (1979), accounting for distance of fall, temperature, relative humidity, and drop size, estimated that rain samples taken during the summers of 1974, 1975, and 1976 had enrichments of 0 to 10.2 ‰, with most of the enrichment occurring during the first part of the storm. High rainfall intensities (greater than 10 mm per hour) are considered to be more representative of the cloud base (Bolin 1959; cited in Yurtsever and Gat 1981). This enrichment is not expected to significantly affect Edmonton precipitation as summer rains for the

Edmonton region are characteristically high in intensity, with July having the greatest intensities: 74% of the five-minute maxima and 56% of the 15-minute maxima are greater than 12 mm h^{-1} (Verma and Toogood 1969).

Correlation of climatic parameters that most likely represent the effects of raindrop enrichment, amount of precipitation and minimum relative humidity, do not increase the correlation coefficient or decrease the standard error unless sequential daily rain events are used (Table 3.6). Sequential rain events would more likely originate from the same water vapour. The use of precipitation amounts does not always result in a large improvement in correlation; however, with one series of data from June 20 to the July 8 the addition of precipitation amount resulted in the correlation coefficient increasing from 0.706 to 0.995 (Table 3.6). Rains during this period were light, ranging from 1.2 to 13.5 mm, and relative humidities low, less than 59%. . The negative rainfall coefficient for this data set agrees with the 'amount effect' (Yurtsever and Gat 1981), the smaller the rainfall the larger the isotope value due to enrichment. A positive coefficient for relative humidity is opposite to that expected for raindrop evaporation, where lower relative humidities mean higher evaporation rates. Relative humidity could also account for some of the effects of isotopic exchange between water vapour and the raindrop.

Table 3.6. Multiple correlation coefficients of sequential rain events.

Dates	Regression Equation Coefficients							
	n	Intercept	T	P	RH	r^2	Probability	SE
1970: July	5	-24.46	+0.655			0.934	0.020	0.825
	5	-23.43	+0.637	-0.044		0.966	0.007	0.242
	5	-32.40	+0.827		+0.072	0.998	0.003	0.163
1971: June	8	-22.37	+0.277			0.721	0.044	0.611
	8	-23.01	+0.311	+0.033		0.747	0.129	0.641
	8	-24.65	+0.357		+0.020	0.853	0.039	0.504
1974: July	5	-23.08	+0.494			0.819	0.089	2.261
1975: June	5	-32.06	+1.057			0.918	0.028	2.899
1986: April	6	-18.74	+0.661			0.910	0.012	1.471
	6	-18.52	+0.648	-0.047		0.912	0.070	1.687
1986: July	7	-37.63	+1.296			0.729	0.063	2.587
	7	-36.83	+1.263	-0.019		0.732	0.216	2.879
1987: June	6	-31.39	+1.275			0.706	0.116	3.442
	6	-8.25	+0.179	-0.954		0.995	0.001	0.564
1987: July	7	-35.05	+1.345			0.884	0.008	3.124
	7	-33.58	+1.128	+0.128		0.952	0.009	2.280
	7	-55.60	+1.547		+0.311	0.974	0.003	1.692

T - Temperature; P-Precipitation; RH-Relative Humidity; SE-Standard Deviation

3.4 SUMMARY AND CONCLUSIONS.

Characterization of meteoric isotope values and their seasonal and daily variation is necessary if the abundances and fluctuations of isotopes in soil and shallow groundwater systems are to be correctly interpreted. Due to hydrodynamic dispersion, diffusion, and high variation in permeability within any soil profile, soil waters can represent precipitation events for a number of years. Here predictive equations would prove useful in the estimation of past meteoric isotopic concentrations. The purpose of this study was to characterize Edmonton meteoric isotope values and develop predictive equations for use in soil and groundwater studies. This involved three aspects: measurement of the meteoric deuterium and oxygen-18 relationship, determination of amplitude and frequency of fluctuation in isotopic values within and between seasons, and the establishment of predictive equations.

The Edmonton meteoric water line for the 1986 to 1987 data is:

$$\delta D = 7.3\ \delta^{18}O - 6.9$$

which is different from that measured by the IAEA in Edmonton during 1961 to 1966:

$$\delta D = 7.7\ \delta^{18}O - 1.0$$

The shift in both slope and intercept is larger than that accounted for by the standard deviations of the regression coefficients. Reduction in slope and intercept are characteristic of precipitation having undergone slightly more alteration by evaporation or precipitation formed more from re-evaporated precipitation.

The average $\delta^{18}O$ value for winter precipitation is -25.9 ‰ (± 4.0) as calculated from monthly values collected by the IAEA. Individual winter events had values within the range of the IAEA data. Oxygen-18 values for individual summer precipitation events are more variable and overall slightly more depleted (-15.3 ± 4.2‰) relative to the average $\delta^{18}O$ for summer using IAEA data (-13.8 ± 1.7‰). The slight enrichment of the IAEA summer data could be due to the inclusion of all precipitation, whereas the individual events were largely made up of precipitation amounts greater than 5 mm. Summer values showed relatively large variations from year to year. In part this could be due to small sample sizes, but it also appears to be due to differing climatic conditions.

Predictive equations for summer values using air temperature exhibited a wide variation with slopes ranging from 0.17 to 1.38. This variation between summers, not accounted for by air temperature, could be due to the different origins of water vapour and also raindrop evaporation and exchange. Much of this variation is likely

due to origins of water vapour as evidenced by snow isotope values which can vary by more than 10 $\delta^{18}O$ ‰ for equivalent temperatures (ie snow is not affected by evaporation and exchange during fall). Possibly a 3 to 4 ‰ shift in oxygen-18 value is due to raindrop evaporation during warm dry conditions. Sequential rainfall events exhibit much smaller standard errors from correlated air temperatures than seasonal correlation values for any one summer do. The inclusion of rainfall amount and minimum relative humidity in multiple correlations with air temperature results in an increase in the correlation coefficient and a decrease in the standard error.

For general (seasonal) prediction of isotope values from climatic data for the Edmonton region, the equations given in Table 3.4 should be used. It is recommended though that due to the high variability shown in summer rains, both within and between years, that daily precipitation be collected and analyzed. Missing data from within seasonal data sets can be more accurately predicted for rainfalls greater than 5 mm, and if sequential rainfalls are used. If predictive equations sufficiently accurate for daily events are to be developed more information about the origins of atmospheric water vapour is required.

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4. THE RELATIVE CONTRIBUTION OF SEASONAL PRECIPITATION TO SOIL AND GROUNDWATER

4.0 INTRODUCTION

Although the amount of snowmelt contribution to soil water may be measured by monitoring overwinter changes in soil moisture content (Staple et al. 1960; Granger et al. 1984) little is known concerning its mixing with resident soil water or its contribution to groundwater. The environmental isotopes, deuterium and oxygen-18, offer this ability.

Seasonal differences in deuterium and oxygen-18 abundances have been used to trace the movement of snowmelt water in the soil profile (Saxena 1984; Bengtsson et al. 1987) and to discern snowmelt runoff from groundwater discharge in streamwater (Krouse et al. 1978; Herrmann et al. 1978; Hooper and Shoemaker 1987; Lawrence 1987; Obradrovic and Sklash 1987). The effects of heavy isotope enrichment and fractionation caused by evaporation can also be used to aid in separation of water from different sources (Gat and Tzur 1967; Gat 1971), to estimate the degree of evaporation (Allison 1982; Barnes and Allison 1983), or to separate the effects of evaporation from transpiration (Allison and Hughes 1983; Barnes and Allison 1983).

Edmonton represents an excellent site for the study of the seasonal contribution of water to the hydrologic system as Edmonton has the second greatest seasonal isotopic variation of 93 global IAEA (International Atomic Energy Agency) sites monitored for meteoric isotope abundances since 1961 (Yurtsever and Gat 1981). Data collected by IAEA (IAEA, 1969, 1970, 1971) has the average winter and summer $\delta^{18}O$ values of precipitation differing by 12 ‰. Other isotopic studies in the Edmonton region have concentrated upon summer rains (Linton 1972; Sandu 1978; Ellis 1979) or upon

origins of atmospheric water vapor (Hage et al. 1975), but none have considered the use of large seasonal differences in isotope abundances to separate snowmelt and rain water in the soil and groundwater systems.

The objective of the study was to determine the amount of snow water contribution to the soil water and groundwater systems. The procedure for determination of the snow water contribution involved the following steps:

1. Measurement of meteoric water parameters: rain and snowfall amounts, air temperature, and the stable isotopes, $\delta^{18}O$ and δD .
2. Measurement of soil water and groundwater parameters and their seasonal fluctuations: soil moisture content, depth to water table, and isotopic ($\delta^{18}O$, δD) abundances.
3. Determination of original isotopic composition (unaltered by evaporation) of soil water and groundwater.
4. Determination of snow water contribution to soil water and groundwater.

4.2 THEORETICAL DETERMINATION OF ISOTOPIC FRACTIONATION DUE TO EVAPORATION.

Evaporation results in the enrichment of the heavier isotopes in the remaining liquid. As deuterium water molecules are different in mass from those of ordinary water and oxygen-18 water, they thus fractionate at different rates and the $\delta^{18}O$-δD relationship of evaporating waters will shift from that of the meteoric water line. This enrichment and alteration of ratios takes place for surface water (Gat 1971) and for soil water (Barnes and Allison 1983). To correct for the effect of evaporation upon soil water, the $\delta^{18}O$-δD lines for both meteoric and evaporated water must be known. These lines may be determined by measurement of the appropriate water or with theoretical calculations. The point where these two lines intersect provides an estimate of the original isotope composition of the input water into the soil.

Evaporation from water bodies commonly results in $\delta^{18}O$-δD slopes of between four and six (Craig et al. 1963; Gat 1971). For dry soils, the evaporation rate becomes limited by the length and tortuosity of soil pores through which liquid water and vapor must pass. As a result, the $\delta^{18}O$-δD slope for soil water can be lower than that of open water, with slopes for soil water ranging as low as 2 or 3 (Dincer et al. 1974; Allison 1982). These low slopes only occur for very dry soils where moisture transport to the evaporation front becomes dominated by vapour rather then liquid movement. For moist situations where liquid transport dominates the evaporation process in the soil, the $\delta^{18}O$-δD slopes are assumed to approximate that of open water bodies (4 to 5). The $\delta^{18}O$-δD slope for soil water undergoing evaporation is directly proportional to air humidity and inversely proportional to the length of the evaporative period (Allison 1982).

The $\delta^{18}O$-δD relationship for soil water may be theoretically determined using an equation developed by Gat and Tzur (1967) which describes the isotopic fractionation resulting from evaporation of ponded water while infiltrating into the soil. The isotopic abundance of the infiltrating water (δs), as affected by the evaporation rate (E), is a function of the air temperature (T), relative humidity (h), the isotopic abundance of the input water (δi), the infiltration rate (I), and the fractionation (ε^*) and kinetic excess separation ($\Delta\varepsilon$) factors:

$$\delta s = \frac{(h\varepsilon^* - \varepsilon)}{h} f^u + \delta i; \text{ where } u = (h - \varepsilon)/(1 - h + \Delta\varepsilon)(1 + I/E) \qquad [4.1]$$

where: ε = enrichment of the liquid if compared with the gas phase
= $\varepsilon^* + \Delta\varepsilon$, where
ε^* = fraction factor = Rc/Rv; where
R = heavy isotope abundance/light isotope abundance
$\Delta\varepsilon$ = kinetic excess separation factor

The fractionation factor, ε^*, represents the unit separation factor at equilibrium and is a function of temperature. Values for ε^* used in this paper are those given by Majoube (1971). The kinetic excess separation factor, $\Delta\varepsilon$, which is the difference between the effective and thermodynamic separation factor can be best approximated for ^{18}O by (Gat 1980):

$$\Delta\varepsilon = 15(1 - h)‰ \qquad [4.2]$$

and for D by:

$$\Delta\varepsilon = 13(1 - h)‰ \qquad [4.3]$$

4.3 MATERIALS AND METHODS

4.3.1 Determination of Snow Contribution to Soil Water and Groundwater.

Annual precipitation may be divided into two components, snow and rain, based not only upon physical state (solid and liquid) or seasonal occurrence, but also upon their isotopic abundances. It is both of these features that makes the environmental isotopes useful for tracer studies. Measured isotope abundances may be used to calculate the proportions of snow (winter precipitation; November thru March) and rain (April thru October) that compose soil water and groundwater. To accomplish this for soil water the isotopic shift due to evaporation must be corrected for. As groundwaters are assumed to be composed of soil water, altered by evaporation, and precipitation, not altered by evaporation, it is necessary to first calculate these proportions, then to correct for evaporation, and finally to calculate the snow contribution.

Groundwater is assumed to be composed of soil water and precipitation based upon the following reasoning. Some soil water will be in poorly accessible pores and may be considered as relatively immobile when compared to other soil water. The proportion

of immobile water increases with increasing clay content and aggregate size (Bengtsson et al. 1987). Immobile water is not totally 'immobile' but only relatively so. The terms 'slow'- and 'fast'-draining water are used in this paper to denote immobile and mobile water. 'Fast'-draining water occur from heavy rainfall or snowmelt events which cause near saturated surface conditions resulting in water flow down macropores, cracks, and root channels directly to the water table. This water is usually that which is gravity 'drained' and thus the pores when water filled have matric potentials approximately -33 kPa and larger. Slow-draining water is that soil water that has resided long enough near the surface to be altered by evaporation and then moves to the water table through the soil matrix by mixing with gravity drained water.. Groundwater is thus assumed to be composed of two waters with distinctive $\delta^{18}O$-δD relationships, that of unaltered precipitation (Eq. [4.4]) and that of soil water having undergone evaporation (Eq. [4.5]):

$$D_p = m_p\,{}^{18}O_p - b_p \qquad [4.4]$$

where; D_p is the δD (‰) of precipitated water,
$^{18}O_p$ is the $\delta^{18}O$ (‰) of precipitated water,
m_p is the slope of the relationship between δD and $\delta^{18}O$ of precipitated water, and
b_p is the intercept of the relationship between δD and $\delta^{18}O$ of precipitated water.

$$D_s = m_s\,{}^{18}O_s - b_s \qquad [4.5]$$

where; D_s is the δD (‰) of soil water,
$^{18}O_s$ is the $\delta^{18}O$ (‰) of soil water,
m_s is the slope of the relationship between δD and $\delta^{18}O$ of soil water, and
b_s is the intercept of the relationship between δD and $\delta^{18}O$ of soil water.

The proportions of soil water and precipitation making up groundwater may be calculated with use of Eqs. [4.6] to [4.11]. The isotopic value of groundwater is

assumed to be made up of proportions of precipitation and soil waters. Either $\delta^{18}O$ (‰) or δD (‰) may be used to represent groundwater:

$$^{18}O_g = x\,^{18}O_p + (1-x)\,^{18}O_s \quad [4.6]$$
$$D_g = xD_p + (1-x)D_s \quad [4.7]$$

where $^{18}O_g$ and D_g represent the isotopic contents (‰) of groundwater,
x represents the precipitation proportion of the groundwater,
$^{18}O_p$ and D_p represent the isotopic contents (‰) of precipitation,
1-x represents the soil water proportion of groundwater, and
$^{18}O_s$ and D_s represent the isotopic contents (‰) of soil water.

As the isotopic concentrations of soil water and groundwater may be measured equations [4.6] and [4.7] have two unknowns in each equation, the isotope concentration of the precipitation ($^{18}O_p$ or D_p) and the proportion of precipitation making up the groundwater (x). By using both equations along with Eq. [4.4] to replace D_p with $^{18}O_p$ the unknowns may be solved for. Rearranging Eqs. [4.6] and [4.7] so as to solve for $^{18}O_p$ and D_p:

$$^{18}O_p = (^{18}O_g - {}^{18}O_s + x\,^{18}O_s)/x \quad [4.8]$$
$$D_p = (D_g - D_s + xD_s)/x \quad [4.9]$$

Replacing D_p in Eq. [4.9] with $^{18}O_p$ from Eq. [4.4] and solving for $^{18}O_p$:

$$^{18}O_p = ((D_g - D_s + xD_s)/ xm_p) + xb_p/xm_p \quad [4.10]$$

As Eq. [4.8] is equivalent to [4.10] they may be combined thus negating $^{18}O_p$ and enabling x to be solved for:

$$x = (D_g - D_s - m_p\,^{18}O_g + m_p\,^{18}O_s)/(m_p\,^{18}O_s - D_s + b_p) \quad [4.11]$$

Now that x is known it may be inserted back into Eqs. [4.8] and [4.9] to solve for $^{18}O_p$ and D_p.

Calculation of the proportions of snow and rain precipitation making up soil water and groundwater may be done through use of equations [4.12] through [4.14]. The first step is to correct for the amount of isotopic shift due to evaporation. For soil water this shift may be estimated by calculating where the meteoric water line (Eq. [4.4]) and the δD-$\delta^{18}O$ relationship of the soil waters (Eq. [4.5]) intercept:

$$^{18}O_{si} = (b_p - b_s) / (m_p - m_s) \qquad [4.12]$$

where $^{18}O_{si}$ is the $\delta^{18}O$ (‰) value at the interception point.

The value of δD at the interception point (D_{si}) may then be calculated by inserting $^{18}O_{si}$ into either of the δD-$\delta^{18}O$ equations [4.4] or [4.5]. If the average seasonal isotope concentrations of snow or winter precipitation (November 1 to March 31) and rain (April 1 to October 31) are known, by measurements, then with D_{si} and $^{18}O_{si}$ it is possible to calculate the proportion of snow precipitation (x) making up soil water:

$$x = (^{18}O_{si} - ^{18}O_r) / (^{18}O_w - ^{18}O_r) \qquad [4.13]$$

where $^{18}O_r$ is the $\delta^{18}O$ (‰) of rain and $^{18}O_w$ is the $\delta^{18}O$ (‰) of snow.

Now that all the above information from Eqs. [4.6] through [4.13] is known it is possible to calculate the proportion of snow in groundwater. The proportion of snow in the precipitation portion of groundwater is calculated using an equation similar to [4.13] and the total proportion of snow in groundwater (x_w) is calculated by multiplying the various components together:

$$x_w = x_p \cdot x_g + x_s \cdot (1 - x_g) \qquad [4.14]$$

where: x_w is the total proportion of snow making up groundwater;
x_g is the snow proportion in the precipitation portion of groundwater;
x_p is the precipitation portion of groundwater (x in Eq. [4.11]);
x_s is the snow proportion of soil water (x in Eq. [4.13]); and
$1-x_g$ is the soil water proportion of groundwater.

the evaporative shift in isotopes in the groundwater is from the soil water it is possible to account for this shift and to calculate the original isotope concentrations of the groundwater on the meteoric water line:

$$^{18}O_{gi} = (1-x_g).^{18}O_{si} + x_g.^{18}O_p \qquad [4.15]$$

where: $^{18}O_{gi}$ is the $\delta^{18}O$ (‰) value of the groundwater on the meteoric waterline (the original value before the effect of evaporation), $^{18}O_{si}$ is the intercept of the soil water and the meteoric water $\delta^{18}O$-δD lines (Eq. [4.12]), and $^{18}O_p$ is the $\delta^{18}O$ (‰) value of the precipitation portion of groundwater (Eq. [4.8]).

4.3.2 Site Description

The study site was located approximately 15 km south of the Edmonton city center at the Ellerslie Research Station. The climate of the Edmonton area is between dry and moist subhumid, characterized by relatively warm summers and cold winters. The average annual temperature is 3°C, with January the coldest month at -14°C and July the warmest month at 16°C. The mean annual precipitation is 440 mm, with 60% of it occurring during May through August and with July having the greatest amount (85 mm). The mean annual winter precipitation (November through March) is 89 mm.

A large portion of the region, within which the study site is located, is under mixed grain cultivation, with barley being the common grain crop. The local topography is gently rolling to rolling glacial lacustrine over morainal plain. The study site is located at the toe to mid-slope position of a 1-2% west slope. Soils in the well-drained to moderately well-drained positions of the Ellerslie Research Station have been described as Eluviated Black Chernozems. Soil texture is clay to clay loam in the top meter with sand lenses occasionally occurring at depths from 0.6 m to 4.0 m. The water table at the study site generally fluctuates between 3 and 4.5 m in depth.

4.3.3 Sampling and Analysis.

Soil samples for isotope analysis were taken from a summerfallow plot and a barley plot that had standing stubble during winter. The plots were 23 m wide by 45 m long and were part of a larger study that monitored the effects of fall cultivation upon snowmelt infiltration for two winters, 1985-86 and 1986-87. Summerfallow, having no overwinter stubble, retains less snow with the consequence of less snowmelt infiltration. Under summerfallow no crops are planted and the soil is tilled to prevent weed growth; therefore the soil does not lose water to transpiration. However, evaporative loses may be higher due to lack of a plant canopy.

When possible, samples were taken to a depth of 4 m within each of the two treatments (Table 4.1). To obtain estimates of site variability replicate samples were occasionally taken to a depth of 1 m. Soil samples were taken approximately every 4 months from March 1986 to July 1987 (Table 4.1). On the first sample date (29 March 1986) samples were taken at 15 cm intervals for the first 1.05 m and at 30 cm intervals thereafter. Samples taken during all other dates were taken at 10 cm intervals for the first 1.0 m then at 20 cm intervals after that. Samples for the top 30 cm were obtained by shovel. Samples beneath 30 cm were taken with an Oakfield sampler. Cores obtained with the sampler were scraped to remove contaminating soil particles and soil from the entire interval was bagged. To obtain sufficient sample for dry conditions in the top meter, two core holes were used. Not all samples were analyzed for isotopes.

Table 4.1. Soil water sampling for stable isotope determination.

Date	Plot treatment	Sites	Depth of samples (m)
29/3/86	standing stubble	3	1.95, 2.25, 3.75
	standing stubble	0	
22/7/86	barley	3	1.00, 1.00, 4.00
	summerfallow	3	1.00, 1.60, 4.40
04/11/86	barley stubble	1	3.60
	summerfallow	0	
14/4/87	barley stubble	2	1.00, 4.00
	summerfallow	2	1.00, 4.00
21/7/87	barley	1	4.20
	barley	1	2.20
21/10/87	barley chiselled	1	2.60
	barley stubble	1	4.00

Rain and snow samples, representative of daily precipitation events, were taken between January 1986 and October 1987 and analyzed for relative oxygen-18 and deuterium abundances. Rain samples were collected using a standard rain gauge located 0.5 km south of the University of Alberta campus. Snow samples for individual events were obtained from a raised surface to prevent accumulation of drifting 'old snow', cleaned of previous snowfall. Some snow samples were taken from accumulated snow at Ellerslie using a snow core. As these samples are of the entire snow profile they cannot be dated to a specific event. Other relevant meteorological information were obtained from the Ellerslie meteorological station.

Water was extracted from the samples with azeotropic distillation using toluene (Allison and Hughes 1983; Hendry 1983). In this method the soil sample is suspended in a cellulose thimble above boiling toluene in a distillation apparatus. The toluene boiling at 110°C turns the soil water into vapour. Both the toluene and water vapours

are condensed and accumulate in a side trap. Toluene being lighter than water and virtually immiscible floats on water. This method has an accuracy of water extraction of ±0.1 ml (Pande 1974). Residual toluene in the collected waters (≈ 100 ppm) was removed with molten wax in stoppered glass test tubes.

Both deuterium and oxygen-18 were analyzed using facilities at the University of Alberta. For isotopic analysis of oxygen-18, 5 ml of water were equilibrated with CO_2 at 25°C for 24 hours followed by isotopic analysis of the CO_2 (Epstein and Mayeda 1953; Gonfiantini 1981). Analysis of deuterium involved reduction of 15 to 20 µL of sample in an evacuated tube with 0.5 g to 0.7 g of Analar zinc shot heated to 450 °C (Coleman et al. 1982). This produced hydrogen gas which was measured by a VG Micromass® 602C. Concentrations of oxygen-18 were determined using a VG Micromass® 602D. All isotope measurements are reported relative to Standard Mean Ocean Water (SMOW).

Overwinter volumetric soil moisture content was measured on the study plot with a neutron moisture meter. Water wells to depths of 3 m and 5 m were located approximately 20 m outside of the study plots.

The $\delta^{18}O$-δD relationship for soil waters was determined both by measurement and by theory using Eq. [4.1]. The current study assumes soil water evaporation is not limited by soil properties and approximates that of a free water surface.

4.4 RESULTS AND DISCUSSION

4.4.1 Meteoric Waters.

The first fall and winter (1985 to 1986) of the study period had near normal temperatures and precipitation (Table 4.2); however, the fall of 1986 had about twice normal precipitation while the second winter (1986 to 1987) was warmer and drier than usual. Total precipitation during the first and second winters of the study was 81 and 50 mm respectively.

Table 4.2. Climatic summary of study period, Fall 1985 to Summer 1987.

	1985 to 1986		1986 to 1987		long term mean†	
Season	T (°C)	P (mm)	T (°C)	P (mm)	T (°C)	P (mm)
Sept. - Oct.‡	5.4	65.4	7.2	109.4	7.4	57.5
Nov. - Mar.	-8.6	80.8	-6.6	49.8	-9.9	89.0
Apr. - May	7.5	84.7	9.5	80.5	6.7	64.9
Jun. - Aug.	14.9	222.5	15.2	235.6	15.0	228.9

† normals are from 24 years of meteorological records for Ellerslie Research Station.
‡ inclusive.

Midwinter melts occurred during both winters, with melts during the first winter being more severe and resulting in almost total reduction of the snowpack. Springmelt for the both years began in the last half of March and finished by the first week in April.

Oxygen-18 values of rain and snow (relative to SMOW) displayed a wider range than that shown by the monthly data (Fig. 4.1) collected at the Edmonton Municipal Airport during 1961 to 1965 by the International Atomic Energy Association (IAEA 1969, 1970, 1971). This is partly to be expected due to the monthly averaging effect of the IAEA data; however, the summer data for 1986 and 1987 had lower and higher isotope

values than those collected by previous studies on Edmonton summer rainfalls (Linton 1972; Ellis 1979). These extreme isotope values are due to colder and warmer than normal monthly temperatures encountered in the summers of 1986 and 1987. July 1986 and August 1987 had the second coldest monthly average temperatures for these months on record for Ellerslie since 1964. The high isotope values during the summer of 1987 are related to the high temperatures in June 1987, which had the highest monthly average for June since 1964.

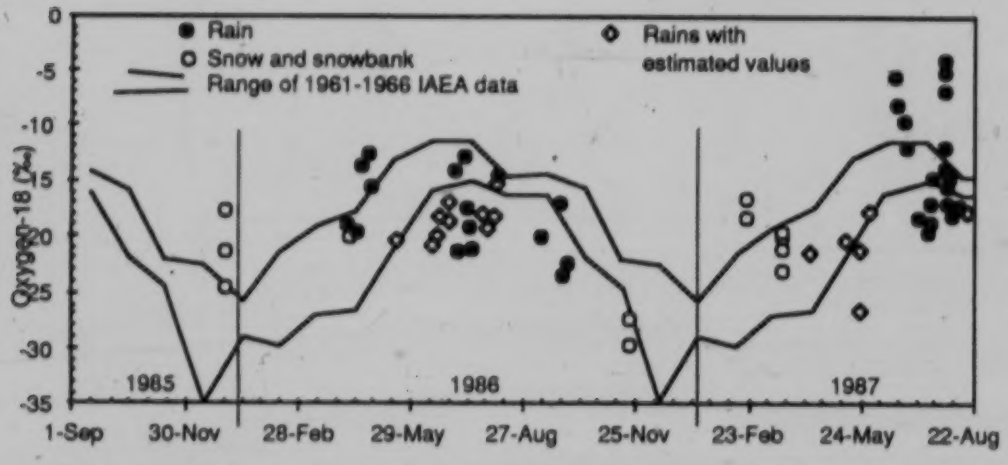


Fig. 4.1. Oxygen-18 values of meteoric waters for 1985 to 1987.

For 1986 and 1987 rain events that were not sampled for isotopes, temperature-$\delta^{18}O$ correlation equations established from values measured in the same season were used for precipitation events greater than 5 mm (Table 4.3)

Table 4.3. Multiple correlation coefficients of sequential rain events.

Dates	n	Regression Equation Coefficients Intercept	T	P	RH	r^2	Prob.	SE
April 1986	6	-18.74	+0.661			0.914	0.012	1.467
July 1986	7	-37.63	+1.296			0.732	0.063	2.593
June 1987	6	-8.25	+0.179	-0.954		0.988	0.001	0.561
July 1987	7	-55.60	1.547		+0.311	0.966	0.003	1.685

T = average daily air temperature (°C); Prob. = probability of a larger F value;
P = daily precipitation (mm); RH = daily minimum relative humidity (%);
SE = standard error

For years in which no isotope data are available regression equations, formed from Edmonton area isotope data (IAEA 1969, 1970, 1971; Linton 1972; Hage et al. 1975; Sandu 1978; Ellis 1979; and this study), were used (Table 4.4).

Table 4.4. Correlation values for $\delta^{18}O$ and temperature.

Temperatures Sample dates	n	Regression equation intercept		slope		r^2	prob. >F	Standard error
T ≤ 0°C	34	-20.55	(1.01)	0.457	(0.07)	0.747	0.000	2.96
T >-5°C, ≤15°C	50	-19.56	(1.05)	0.208	(0.09)	0.303	0.034	2.83
T > 15°C	35	-34.81	(4.26)	1.225	(0.24)	0.664	0.000	3.18

For temperatures below 0°C daily or monthly mean temperatures may be used with the equation in Table 4.4, whereas for temperatures above 0°C the equations in Table 4.4 should be used with daily mean temperatures for daily rains greater than 5 mm. An assumption was made that any daily rains less than 5 mm would likely be evaporated at the soil surface and thus not contribute to the soil or groundwaters (Gat 1971). All $\delta^{18}O$ values for meteoric waters may be converted to δD values with the following

$\delta^{18}O$-δD equation established from the 1986 to 1987 precipitation samples collected in Edmonton:

$$\delta D\ (‰) = 7.29\ (\pm 0.16)\ \delta^{18}O\ (‰) - 6.88\ (\pm 2.67) \qquad [4.16]$$

where; the numbers in brackets are standard deviations

n = 52,

probability of a larger F = 0.000,

coefficient of correlation = 0.989,

standard error = 6.16

Measured seasonal isotope values (weighted to precipitation amount) for the study period are compared with estimated isotope values from Table 4.4 to evaluate the accuracy of these equations (Table 4.5).

Table 4.5. Total rain and snow amounts and isotope values, 1985-87.

		Precipitation (mm)		$\delta^{18}O$ ‰ (±error estimate)	
Season	year	Seasonal total	measured†	measured§	estimated‡
winter	85-86	80.8	.. snowbank* ..	-21.2 (3.4)	-23.4 (0.5)
spring	86	84.7	43.3	-17.9 (2.4)	-18.5 (0.4)
summer	86	222.5	99.4	-18.1 (1.9)	-16.2 (2.4)
fall	86	109.4	87.8	-20.5 (3.8)	-18.2 (0.6)
winter	86-87	49.8	.. snowbank* ..	-21.6 (4.3)	-23.6 (0.5)
spring	87	80.5	-	-	-18.5 (0.5)
summer	87	235.6	150.0	-13.0 (4.2)	-14.5 (3.1)
fall	87	62.4	53.6	-18.3 (n/a)	-17.3 (0.9)

† precipitation measured for $\delta^{18}O$.

§ isotope values are weighted (to precipitation amount) averages. Error estimate is standard deviation of weighted amounts.

‡ $\delta^{18}O$ predicted from general equations Eqs. in Table 4.4. Error estimates calculated by varying slope by their standard deviations.

* samples from snowbank, so cannot estimate precipitation measured.

Generally the equations of Table 4.4 estimated seasonal isotope values to within 1 to 2 $\delta^{18}O$ ‰. Although air temperature is the main factor influencing the isotope abundance in precipitation, other factors such as humidity, amount of rainfall, and origin of the

water vapor affect and confound prediction attempts. Differences between the measured and predicted isotope values in Table 4.5 are likely due to the effect of different origins of atmospheric water vapor (Hage et al. 1975).

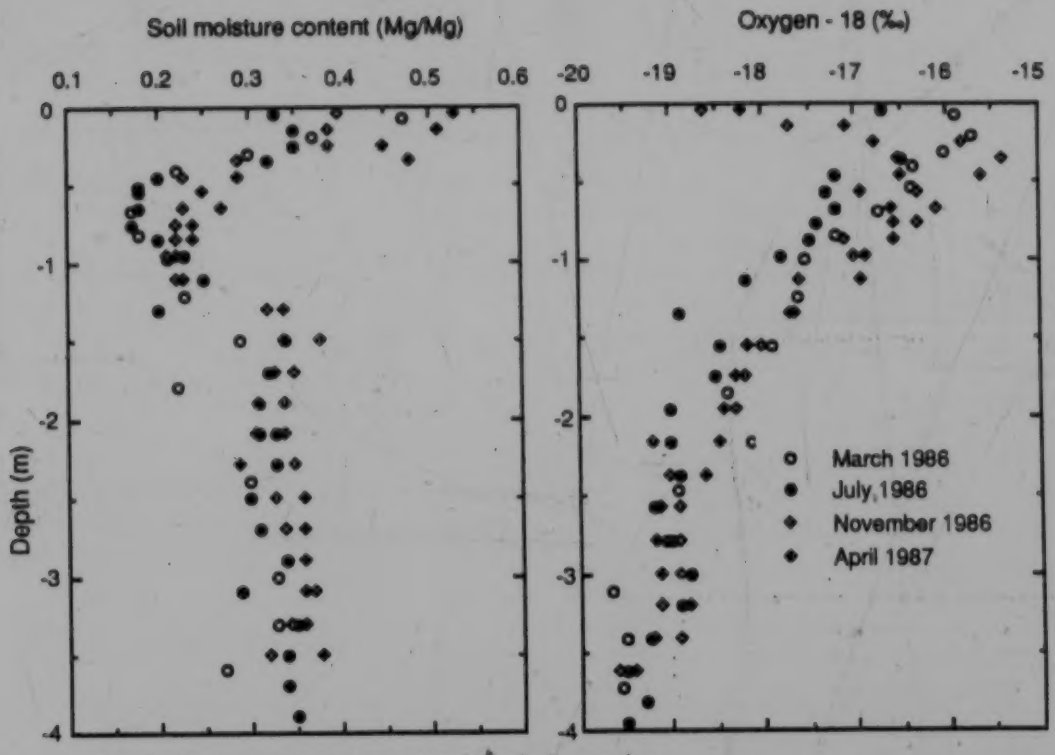
Due to the clay to clay loam texture of the soil at Ellerslie, there is likely a large proportion of 'immobile' water. Thus waters in the root zone and especially the groundwater may be a mixture of several years to several decades of previous meteoric waters. Lacking any other information it is assumed that the water in the soil and groundwater represents the previous five years of climatic data (1981 to 1986). Isotopic values for snow and rain averages for these five years were generated from recorded temperatures with the equations in Table 4.4. The estimated seasonal values are -25.1 (±2.0) ‰ for snow and -16.7 (±1.1) ‰ for rains. The annual $\delta^{18}O$ average, weighted for precipitation, for Ellerslie (1981 to 1986) is estimated at -18.8 ‰. As compared to the IAEA data, collected for the years 1961 to 1966, the winter values are the same, but the IAEA rains are higher, at -15.2 ‰. Some of this could be due to the inclusion of rains less than 5 mm with the IAEA data, which account for 23.7 % of the total rainfall.

4.4.2 Soil Waters and Groundwaters

The isotopic concentration in the profile decreased in value from 0.3 m to 4.0 m and in the standard deviation from the surface down to 4.0 m (Fig. 4.2, Table 4.6). With the exception of the surface 1.0 m the isotope values remained did not vary with time (Fig. 4.2, Table 4.7). The greatest fluctuation in both soil moisture content and isotopes was in the top 0.3 m (Fig 4.2).

During the study period the water table fluctuated between 3.7 and 4.4 m. Discounting any textural changes at 1.3 m it is assumed that the sudden change in moisture content at that depth could represent the upper extent of the capillary fringe. The division between the unsaturated and the saturated zones is considered to be the upper extent of the capillary fringe, at which suction is equal to the air-entry value of the soil (Hillel 1980). The high standard deviations of the top meter could be due to the unsaturated conditions constantly under change due to evaporation, transpiration, and precipitation, whereas below this depth, the higher moisture contents in the capillary fringe would moderate any moisture removals or additions in the root zone.

The differences in the soil water isotope concentrations was not as large as those between precipitation of different seasons or consecutive precipitation events. For example following the 1986 springmelt, the average $\delta^{18}O$ from three sites in the top 35 cm was -15.9 ‰ (n=9) despite an overwinter soil water increase of 0.21 ($m^3\ m^{-3}$) from snow with a $\delta^{18}O$ value of -21.2 ‰. Also heavy rains during a seven day period beginning July 12, 1986 (63 mm with an average $\delta^{18}O$ of -19.7 ±1.6‰) seemed to only affect the soil water isotope values of the top 0.1 m (-17.4 ±0.8 ‰, n=4, sampled on July 22, 1986) whereas the isotope values for the 0.1 to 0.2 m layer remained high (-14.3 ±0.7 ‰, n=4). This lack of response by the isotopes in soil water to those of rainwater and snowmelt water could be due to water from the heavy rains rapidly infiltrating via cracks, channels, and interaggregate spaces and than mixing with the already resident soil water. The soil at Ellerslie has a high amount of clay and this could result in a large amount of relatively immobile waters which would mix with the infiltrating waters rather than being displaced by them.



values in figures are not representative of all samples.

Fig. 4.2. Distribution of soil water content and $\delta^{18}O$ (‰) with depth.

Table 4.6. Isotopic composition of soil waters and groundwaters; by depth.

Conditions (m)	n	$\delta^{18}O$ (±sd)‰	δD (±sd)‰	θ (±sd)m^3 m^{-3}
Both plots all dates*				
0-0.3 m	27	-15.8 (2.32)	-135.2 (7.34)	0.39 (0.063)
0.3-0.6 m	27	-15.8 (0.90)	-132.4 (3.52)	0.25 (0.048)
0.6-0.9 m	22	-16.6 (0.60)	-135.4 (3.44)	0.22 (0.040)
0-0.9 m	76	-16.1 (1.55)	-134.2 (5.32)	0.29 (0.094)
1.0-2.0 m	30	-17.8 (0.90)	-143.7 (4.23)	0.27 (0.055)
2.0-3.0 m	21	-18.8 (0.38)	-149.8 (1.99)	0.32 (0.025)
3.0-4.0 m	14	-19.3 (0.26)	-151.7 (2.03)	0.34 (0.038)

* only samples which had complete analysis; oxygen-18, deuterium, and water content, were included for this summary.

Table 4.7. Isotopic composition of soil waters and groundwaters; by depth, tillage treatment, and sample date.

Conditions (m)	n	$\delta^{18}O$ (±sd)‰	δD (±sd)‰	θ (±sd)$m^3\ m^{-3}$
Barley 0 - 0.3 m				
March 86†	6	-16.1 (1.28)	-135.2 (8.22)	0.42 (0.056)
July 86	6	-14.6 (2.69)	-133.1 (6.89)	0.35 (0.027)
Nov. 86	3	-17.3 (1.42)	-140.2 (6.11)	0.39 (0.008)
April 87	3	-17.4 (0.76)	-139.8 (2.83)	0.52 (0.015)
Barley 0.3-0.6 m				
March 86†	6	-16.0 (0.77)	-131.6 (1.64)	0.24 (0.061)
July 86	6	-16.3 (0.80)	-134.6 (3.59)	0.23 (0.050)
Nov. 86	3	-15.8 (0.44)	-130.9 (1.57)	0.25 (0.030)
April 87	3	-15.5 (2.17)	-133.8 (4.83)	0.28 (0.038)
Barley 0.6-0.9 m				
March 86†	4	-16.6 (0.57)	-137.4 (3.29)	0.18 (0.015)
July 86	3	-17.4 (0.15)	-137.2 (1.87)	0.18 (0.019)
Nov. 86	3	-16.5 (0.54)	-135.3 (1.60)	0.22 (0.002)
April 87	3	-16.6 (0.01)	-132.3 (3.85)	0.25 (0.015)
Barley 1.0-2.0 m				
March 86†	9	-17.8 (0.59)	-144.3 (3.27)	0.25 (0.031)
July 86	5	-18.2 (1.09)	-146.3 (2.41)	0.28 (0.059)
Nov. 86	5	-18.0 (0.29)	-146.8 (2.59)	0.32 (0.061)
April 87	5	N/A	-143.0 (1.73)	0.31 (0.044)
Fallow 0-0.3 m				
July 86	6	-14.5 (3.08)	-131.5 (8.98)	0.35 (0.048)
April 87	3	-17.2 (1.40)	-136.7 (6.25)	0.39 (0.023)
July 87	3	N/A	N/A	0.28 (0.047)
Fallow 0.3-0.6 m				
July 86	6	-15.5 (0.43)	-132.2 (4.91)	0.22 (0.032)
April 87	3	-15.7 (0.73)	-129.7 (1.47)	0.29 (0.047)
July 87	3	N/A	-130.3 (1.06)	0.19 (0.009)
Fallow 0.6-0.9 m				
July 86	6	-16.4 (0.87)	-137.0 (3.13)	0.20 (0.012)
April 87	3	-16.6 (0.22)	-131.1 (0.25)	0.29 (0.013)
July 87	3	N/A	-133.5 (2.72)	0.21 (0.010)
Fallow 1.0-2.0 m				
July 86	6	N/A	-144.0 (2.82)	0.31 (0.039)
April 87	5	N/A	-144.8 (4.14)	0.29 (0.031)
July 87	5	N/A	-142.9 (5.64)	0.28 (0.054)

† samples taken on March 1986 were from the following depth intervals: 0-0.15 m, at 10 cm intervals to 0.45 m, then at 15 cm intervals to 1.05 m, then at 30 cm intervals. Samples taken for all other days were at 10 cm intervals from 0 to 1.0 m than at 20 cm intervals.

N/A not available.

4.4.3 Isotopic enrichment due to evaporation

The isotope values of soil water to the depth sampled, 4 m, have been enriched due to evaporation, as indicated by their δD-$\delta^{18}O$ values which plot to the right of the meteoric values (Fig. 4.3). The amount of shift from the meteoric water line decreases with depth with the water from the root zone (0.1 - 0.9 m) having the greatest shift. Isotope analysis of soil water from the surface 10 cm were not included as these tended to more composed of recently infiltrated precipitation water unaffected by evaporation. Enrichment due to evaporation for the soil waters is also indicated by the soil δD-$\delta^{18}O$ relationship which has a lower slope and δD intercept than the meteoric relationship (Fig. 4.3 and Table 4.8). The lack of a similar shift of groundwater from the meteoric line as compared to the soil waters is indicative that the groundwaters are not composed entirely of water from the soil profile.

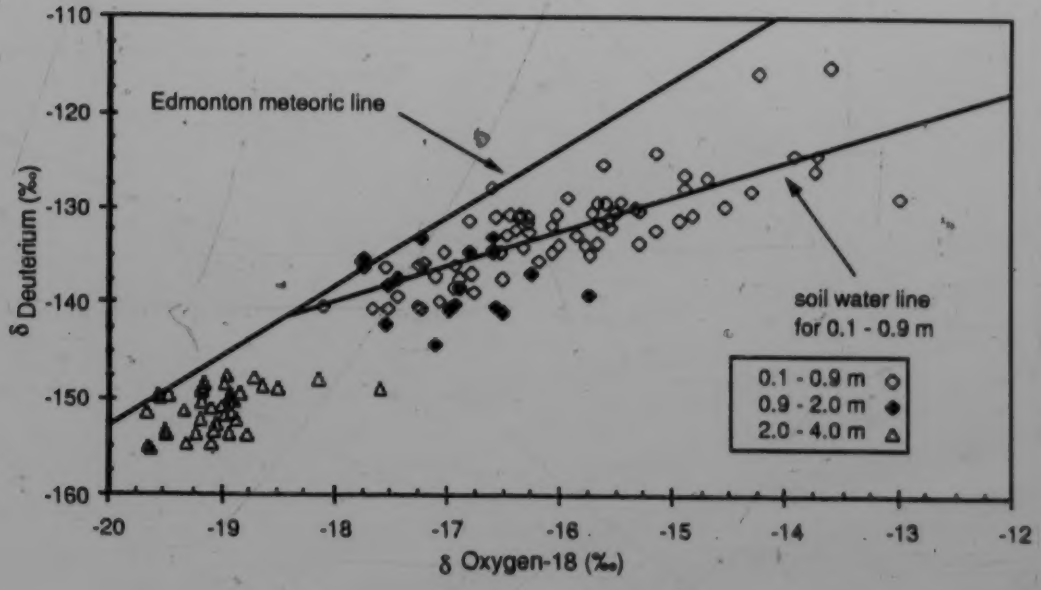


Fig. 4.3. Deuterium and oxygen-18 values for soil water and groundwater

Table 4.8. δD and $\delta^{18}O$ water lines indicating evaporation in soil water.

Condition	slope		δD intercept (‰)		n	r^2	prob. >F	SE
Edmonton precipitation (1986-88)	7.29	(0.15)	-6.88	(2.67)	52	0.989	0.000	6.16
Central Alberta lakes (Swartz 1981)†	4.49	(0.23)	-54.08	(2.14)	12	0.987	0.000	0.44
Barley, 0.1 - 0.9 m								
March 1986	4.45	(0.74)	-61.39	(11.90)	16	0.849	0.000	2.33
July 1986	3.21	(0.47)	-81.69	(7.66)	10	0.906	0.000	2.06
Nov. 1986	4.25	(0.66)	-65.05	(10.81)	8	0.934	0.001	1.34
July 1987	3.82	(1.12)	-68.14	(17.99)	6	0.863	0.027	4.20
Fallow, 0.1 - 0.9 m								
July 1986	4.80	(0.70)	-57.77	(10.89)	15	0.886	0.000	2.76
April 1987	3.76	(0.85)	-69.96	(13.95)	7	0.894	0.007	1.79
Soil water average, both plots, all dates								
0.1 - 0.9 m	3.48	(0.32)	-77.03	(5.17)	72	0.791	0.000	2.44
Theoretical, Eq. [4.1]	3.43	(0.04)‡	-77.49	(2.00)‡				

† unpublished material (Geology 583, University of Alberta).
‡ standard deviations for theoretical representation obtained by varying inputs by 20%.

For the soil δD-$\delta^{18}O$ relationships the lowest slope and intercept were found for the barley plot during July 1986 (Table 4.8). Due to the lack of a plant canopy on the fallow plots, evaporative losses from these plots are expected to be higher than that from the barley plots. Thus it would have been expected that the fallow plots would have had the greatest shift in the δD-$\delta^{18}O$ relationship. Isotopic fractionation is not involved with water uptake by plant roots as transport is liquid and does not involve a change in phase (Zimmermann et al. 1967), unless conditions are very dry resulting in shrinkage of plant roots (Allison and Huges 1983). The most enriched isotope values in soil water should occur at the evaporation front, located within the top 0.1 m. Diffusion into less enriched deeper water would result in a gradual decrease of the heavy isotopes with depth (Zimmermann et al. 1967; Allison 1982; Barnes and Allison 1983), yet the isotope values at 0.3-0.6 m were at least as high as those at 0-0.3 m (Fig. 4.2, Tables 4.6 and 4.7). This higher than expected enrichment midway in the root zone and the higher evaporative effect displayed by the $\delta^{18}O$-δD line of the barley

plots is possibly the result of water removal by plant roots which created a matric gradient thus causing the downward flow of the enriched surface waters (which have undergone evaporation).

Another factor contributing to the greater isotopic enrichment and a greater lowering of the $\delta^{18}O$-δD slope for the barley plot than the fallow is the higher soil water content of the fallow plot. This is assuming that much of the soil water at these depths is made up of precipitation that has directly infiltrated and has thus not been affected by evaporation. This soil water dilutes the waters affected by evaporation. Barley plants results in removal of much of the soil water in the root zone, thus reducing the amount that surface water, affected by evaporation, is diluted (Barnes and Allison 1983).

Low soil water $\delta^{18}O$-δD slopes between 3.2 and 3.8 correspond with the theoretical value of 3.4 (Eq. [4.1]). Higher slopes (> 4) are from wetter conditions such as that following snowmelt in 1986, fallow conditions in July 1986, and fall 1986. An exception occurs for the fallow plot on April 1987 which had a low slope corresponding to drier conditions (Table 4.8). This could have been due to site variability.

The slopes and intercepts in Table 4.8 exhibit a linear relationship if plotted against each other (Fig. 4.4). This is indicative of a mixing line comprised of two different waters; one not altered by evaporation - that of meteoric waters; and one altered by evaporation - that of soil waters. The slope and intercept values for the barley plots on July 1986 and Edmonton precipitation represent the ends of the mixing line. Higher slopes encountered in soil waters are thus due to a mixture of waters; water altered by evaporation and water (precipitation) not altered by evaporation.

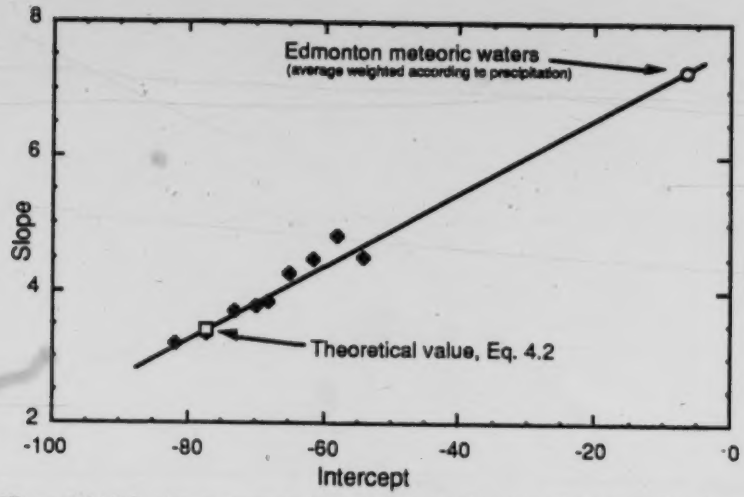


Fig. 4.4. δD and δ18O slope versus intercept values of soil and meteoric water.

The theoretical determination of the effects of evaporation on the isotopes of infiltrating waters (Eq. [4.1]) used the following climatic and soil properties as input parameters:

- the average temperature and humidity for the months May through August, 13.2 °C and 61 %;
- isotope values for precipitation waters was the intersection of the soil water evaporation line for the barley plots on July 1986 with that of the Edmonton meteoric line, (-17.7 ‰, -135.6 ‰),
- evaporation was set at 2 mm per day. The evaporation rate was obtained from soil evaporation losses on a fallow plot during May through August at Ellerslie in 1983 (Maulé and Chanasyk 1987).
- So as to estimate the near-maximum affect of evaporation on soil-water, infiltration was set at low value of 2 mm per day so the water would remain at the surface and may be freely available for evaporation.

The theoretical slope was calculated to be 3.43. A sensitivity analysis of this slope was accomplished by changing the temperature, humidity, and input isotope abundance by 20% or the evaporation or infiltration rates by an order of magnitude. This resulted in

a maximum slope change of 0.04 and a maximum intercept change of 4.4, less than the standard deviation of any of the soil water $\delta^{18}O$-δD lines in Table 4.8.

4.4.4 Snowmelt Composition of soil water and groundwater

The proportion of snowfall to total precipitation for the winters of 1985 to 1986 and for 1986 to 1987 was 22% and 10% respectively; whereas the long-term average is 19%. Neutron probe measurements during these winters on the barley plots indicated that snowmelt increased soil moisture content in the top 0.90 m by 16% and 4% for the first and second winters respectively. By using the δD-$\delta^{18}O$ water lines for the root zone soil water listed in Table 4.8 the soil water may be corrected for evaporation by calculating the intercept of the soil water evaporation δD-$\delta^{18}O$ line with the meteoric water line (Eq. [4.12]) . These corrected values, being the intersection with the meteoric water line, and average seasonal isotope values of precipitation (a $\delta^{18}O$ of -25.1 ‰ for snow and -16.7 ‰ for rains) are then used with Eq. [4.13] for soil water and Eq. [4.14] for groundwater to calculate the amount of snow water making up soil water.

Using the depths and conditions listed in Table 4.8 the calculated amount of snow water in soil water ranges between 11% and 45% (Table 4.9). The error in brackets next to the estimates (Table 4.9) is calculated by varying the winter precipitation isotope value, -25.1 ‰, by its standard deviation, ±2.0 ‰. The higher snow water estimates occurred for $\delta^{18}O$-δD slopes greater than 4, which represent mixtures of soil water subjected to evaporation and unevaporated precipitation. The lower slopes, occurring on July 1986 and July 1987 (barley plots), are more representative of evaporative conditions and provide estimates of snow water composition of 19.1 % and 11.4 %

respectively. These values correspond more closely with the neutron probe measurements.

Table 4.9. Soil isotope values corrected for evaporation and snowmelt composition of soil waters and groundwaters.

Conditions	n	intercept with meteoric line† $\delta^{18}O$	intercept with meteoric line† δD	Snow waters (% ± SD*)
Barley, 0.1 - 0.9 m				
March 1986	16	-19.2	-146.9	29.9 (5.1)
July 1986	12	-18.3	-140.5	19.1 (3.1)
Nov. 1986	8	-19.1	-146.1	28.6 (4.8)
July 1987	6	-17.7	-135.6	11.4 (1.6)
Fallow, 0.1 - 0.9 m				
July 1986	15	-20.5	-156.1	44.9 (7.8)
April 1987	7	-17.9	-137.3	14.2 (2.1)
Soil water, both plots, all dates				
0.1 - 0.9 m	72	-18.4	-141.1	20.3 (3.3)
Theoretical [Eq. 4.1]		-18.3	-140.5	18.6 (6.8)
Groundwater, both plots, all dates				
3.0 - 4.0 m	16	-20.4	-155.3	43.6 (7.6)

† isotope value corrected for evaporation occurs at the intercept of the soil water and meteoric lines.

* SD Standard Deviation estimated by using standard deviation of $\delta^{18}O$ average seasonal winter, ±2.2.

Groundwater is assumed to be not only composed of snow and rain waters, but also soil water that has undergone evaporation and 'direct' precipitation (that has not been evaporated). In order to the amount of snow water in groundwater it is thus necessary to account for these other waters. This was done with Eq. [4.13]. The estimated snow water contribution to the groundwater is 43.6% (Table 4.9). About 56% of the groundwater is composed of 'fast'-draining soil water that did not reside long enough in the soil zone to be altered by evaporation, while the remainder, 44%, is composed of soil water. The $\delta^{18}O$ and δD values of the precipitation portion of the groundwater (Eqs. [4.8] and [4.9]) are -21.9 ‰ and -166.5 ‰ respectively. The proportion of snow water in the precipitation portion of the groundwater (calculated with a variation of

Eq. [4.13]) is 61.8% (±21.8). A possible explanation for the higher snow water contribution in groundwater is that snowmelt runoff collected in local shallow depressions resulting in point recharge to the groundwater table and subsequent 'mounding'. The groundwater table fell during the winters (1985 to 1986: -60 cm; 1986 to 1987: -18.5 cm) and rose during the summers (1986: +63 cm; 1987: +145 cm). As snowmelt runoff accumulates in shallow depressions, groundwater mounding may develop resulting in lateral flow to other portions of the landscape. Such runoff and ponding is observed during spring snowmelt due to a rate of melt in excess of the rate of infiltration of the frozen soil.

This accumulation of snowmelt waters in small local depressions and resulting point source recharge in combination with some direct recharge of rainwaters could explain both the greater isotopic depletion of the groundwaters and the relative lack of evaporative alteration as compared to the soil waters. Another possible source of high snow water contribution to the water table could be from nearby woodlands, which also occupy the lower portions of the landscape. They tend to trap windblown snow, have a more shallow depth of freezing, and have a slower melt rate than the adjacent agricultural lands.

4.5 SUMMARY AND CONCLUSIONS

Oxygen-18 values of meteoric waters may differ by about 8 ‰ between seasonal averages of rain and snow and by as much as 10 to 15 ‰ for individual events, but due to the nature of the soils of the study site these fluctuations are not reflected in soil water. This could be due to the large amounts of relatively immobile water caused by the clay to clay loam texture, along with preferential water flow routes caused by structural cracks, soil fauna, and root channels during near-saturated conditions of

heavy rain fall. Infiltrating waters do not displace the already resident soil water but mix with them. This makes it difficult to discern the result of any specific meteoric or even seasonal event in the soil profile. It is possible, however, to obtain a general estimate of the contribution of snowmelt to soil water and groundwater with the use of isotopes. Using measured isotope values of seasonal precipitation and of soil water the contribution of winter precipitation to soil water (0.1 m to 0.9 m) was calculated to be between 19% and 11% for the barley conditions of July 1986 and July 1987. Overwinter soil moisture measurements with the neutron probe indicated that snowmelt accounted for increases in soil moisture of 16% for the 1985 to 1986 winter and 4% for the 1986 to 1987 winter.

Groundwater had a lower isotope concentration than soil water and was also less affected by evaporation. Assuming that groundwater was composed of soil water (that altered by evaporation) and precipitation (that not altered by evaporation), it was possible to calculate the proportions of snow water and rain water. Groundwater thus was estimated to be composed of 44% soil water and 56% precipitation. In terms of seasonal composition groundwater was estimated to be also composed of 44% snow water and 56% rain water (Authors note: these proportions of 44% and 56% are coincidentally the same proportions as that of soil water and precipitation; the calculations have been checked) with about 62% of the precipitation portion of groundwater being due to snow water. The high proportion of unaltered precipitation and snow waters in groundwater is postulated to be to the impermeable conditions of frozen soil and the resulting accumulation of snowmelt runoff in shallow local depressions. If snowmelt did collect in local depressions and infiltrate, there would be little alteration by evaporation due to the cool temperatures of the spring months.

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5. SIMULATION OF SEASONALLY VARYING ONE-DIMENSIONAL SOLUTE TRANSPORT: I. MODEL DESCRIPTION AND SENSITIVITY ANALYSIS.

5.1 INTRODUCTION

Studies of snow on the Canadian prairies have documented snowtrapping and the resulting increases in soil moisture, but very few have quantified what happens to the melt water once in the profile, and more importantly, determined the 'histories' of seasonal soil water additions. Studies by Meyboom (1966) and Maclean (1974) have been the exception. Their research linked fluctuations in the local groundwater system to seasonal climatic trends. The Canadian prairies have distinct seasons; with winters temperatures frequently below -20°C, and summer temperatures occasionally reaching as high as 35°C. It is unimaginable that these extremes would not be reflected in the soil and shallow groundwater systems. The purpose of this, and the next chapter, is to evaluate seasonal trends in moisture flow between the unsaturated root zone and the water table with the use of numerical simulation techniques.

Due to concern with the presence of pollutants and salts in soil water and groundwater, there has been intense interest in modelling the flow and dispersion of solutes in soil and groundwater systems (Frissel et al. 1970; Leistra 1973; Cameron et al. 1979; Bresler et al. 1982). The study of flow and dispersion in porous media is dependent upon the presence of a tracer or pollutant; consequently, studies are usually limited to situations in which the tracer is added and not part of the natural environment. Fortunately the large differences in seasonal abundances in the deuterium and oxygen-18 isotopes of precipitation in Alberta (International Atomic Energy Agency (IAEA) 1969, 1970, 1971; Freeman 1986) make a seasonal study possible.

Environmental isotopes also have the advantages over other tracers of being naturally occurring and behaving like normal water ($H_2{}^{16}O$) in their flow behaviour (Bengtsson et al. 1987).

The objectives of this section are to develop a flow and mixing model for the one-dimensional transport of non-reactive solutes in a soil profile, capable of simulating flow that seasonally changes direction, and to perform a sensitivity analysis of the model input parameters. Flow could vary seasonally: upwards due to evapotranspiration or towards a freezing front and downwards during wet seasons or after springmelt. The sensitivity analysis will evaluate the relative importance of the various input parameters in affecting the output, which is the solute concentration. This model simulates only one-dimensional flow and assumes that there is no additions or removals via lateral flow and also that there is no preferential flow routes that can bypass other flow routes during saturated flow. The following chapter (Chapter 6) applies the model, developed in this chapter, to field data collected at Ellerslie, Alberta.

5.2 MODEL DEVELOPMENT

5.2.1 Introduction

The numerical equations used here are similar to those developed by Bresler (1973); Leistra (1973); van Genuchten and Wierenga (1974); and Bresler et al. (1982). These equations have been validated, mathematically and experimentally, by these authors but not for alternating flow. Analytical solutions for alternating flow would be complex, thus making it difficult, if not impossible, to use them for validation. Currently the only measurements collected are field data used to establish the model and thus data are not available for validation. Model evaluation in this section is accomplished by a sensitivity analysis. A sensitivity analysis involves the evaluation

of relative changes in output by systematically varying input parameters, one at a time, while holding all other input parameters constant (Crowe and Schwartz 1981). This form of evaluation provides an estimate of allowable error in the measurement of input parameter when the model is applied to real situations. The sensitivity analysis also provides a necessary evaluation of nodal spacing for the size of space and time steps.

5.2.2 Diffusion and Hydrodynamic Dispersion

When water of differing solute concentration than that resident move through the soil system, diffusion and dispersion processes result in mixing of the fluids. The diffusion (D_w^{dif}) for the solute in bulk water occurs as a result of a change in solute concentration (dC_w) in bulk water, over a change in distance (dz_w). Hydrodynamic dispersion (D^{disp}) results from differences in flow velocity from point to point and may be treated in a manner analogous to diffusion. Flux for convective transport in the soil water phase is equal to $q\ \theta^{-1}\ C_w$, where q is flux ($m\ s^{-1}$) in soil, and θ is the volumetric moisture content.

Although there is no published information for the diffusion of oxygen-18 water ($H_2{}^{18}O$) in normal water ($H_2{}^{16}O$) the diffusion rate of HD ^{16}O in normal water (1.295×10^{-9} m^2/s at 5 °C) and the self diffusion rate of normal water (1.301×10^{-9} m^2/s at 5°C; Mills 1973) is sufficiently close enough in value that for the purposes of this study all three above diffusion rates were represented by the self diffusion of normal water (Harris and Woolf 1980):

$$D_w^{dif}\ (\times 10^{-9}\ m^2/s) = 0.00026\ T^2 + 0.04328\ T + 1.06844 \qquad [5.1]$$

for the temperature range between 5°C and 45°C.

The diffusion of any tracer in the soil, D_s^{dif} is affected by the porosity (n) of the soil and the volumetric moisture content (θ). Shearer et al. (1973) developed an equation to correct for the effects of tortuosity upon diffusion:

$$D_s^{dif} = (\theta/n)^2 \, (\theta^{4/3}) \, D_w^{dif} \qquad [5.2]$$

where D_w^{dif} is the coefficient of self diffusion of oxygen-18 water.

As both diffusion and dispersion result in spreading of a solute, they are commonly combined into one process, apparent hydrodynamic dispersion, D^{spr}. Diffusion effects become more important at low flux densities, as dispersion being dependent upon flux, becomes smaller (Miller 1983).

No measurements of dispersion were made in this study, and thus literature values were used. Apparent hydrodynamic dispersion (D^{spr}), as a function of velocity of flow, was determined from data collected by Nielsen and Biggar (1962, 1963; cited in Olsen and Kemper 1968) and is described by the following:

$$D^{spr}(m^2/s) = \{-0.64 + 1.12 \ln(100v) - 0.181 [\ln(100v)]^2 + 0.0188 [\ln(100v)]^3\} \times 10^{-9} \qquad [5.3]$$

where v is the velocity in cm/d. Equation [5.3] was used for estimation of the apparent dispersion coefficient in the model. The diffusion of water, Eq. [5.1], was modified for soil properties, moisture and tortuosity, with equation [5.2]. At low flow rates (approximately 1.1×10^{-9} m/s or 0.01 cm/d), the apparent dispersion rate, D^{spr}, is equivalent to diffusion, D_s^{dif}, of Eq.[5.2].

5.3.3 The flow equation.

The effects of diffusion and convection upon solute movement for transient one-dimensional flow, in which the solute is inert with respect to the soil and water phases can be described by the following equation (Bresler 1973; Bresler et al. 1982):

$$\frac{\partial(c\theta)}{\partial t} = \frac{\partial}{\partial z}\left[D(\theta,\frac{q}{\theta})\frac{\partial c}{\partial z} - qc\right] \qquad [5.4]$$

where c is the solute concentration, t is time, and q is moisture flux. One of the main problems with finite difference solutions of the dispersion equation is that considerable 'numerical dispersion' or pseudo-spreading can occur. Numerical dispersion is a result of the 'step interval', Δz, and acts in the same way as physical dispersion. Its value depends on the flow velocity as well as moisture content and the space and time increments (Bresler 1973). Other numerical techniques such as the finite element method (Wang and Anderson 1979) or the probabilistic-deterministic mass transport model (Ahlstrom et al. 1977; Schwartz 1978) avoid the problem of numerical dispersion, but these techniques are more applicable to two and three dimensional flow. For conditions of one-dimensional flow with a regular mesh network and with simple boundary conditions, the finite element and the finite difference methods are equivalent (Bresler et al. 1982). Finite difference methods may be used with the selection of proper correction expressions for numerical dispersion (Chaudhari 1971; Lantz 1971; Bresler 1973; van Genuchten and Wierenga 1974; Bresler et al. 1982).

The following technique for minimizing numerical dispersion is from Bresler et al. (1982) as developed by Chaudhari (1971); Lantz (1971); and Bresler (1973). Numerical dispersion is calculated from first-order finite difference approximations to the first-order time and space derivatives in Eq. [5.4] (i.e., $\partial(c\theta)/\partial t$ and $\partial(qc)/\partial z$); consequently at least second-order approximations of these derivatives

must be considered (Bresler et al. 1982). The following expression is a second-order finite difference equation approximating the partial differential equation for flow of non-interacting solute that eliminates the effect of numerical dispersion. It has been validated by comparison with analytical solutions and with field data (Bresler 1973; Bresler et al. 1982):

$$\frac{\theta_i^{j+1}c_i^{j+1} - \theta_i^j c_i^j}{\Delta t} = \frac{D_{i-1/2}^{j+1/2} - N_{i-1/2}^{j+1/2}}{2(\Delta z)^2}\left(c_{i-1}^{j+1} + c_{i-1}^j - c_i^{j+1} - c_i^j\right)$$

$$- \frac{D_{i+1/2}^{j+1/2} - N_{i+1/2}^{j+1/2}}{2(\Delta z)^2}\left(c_i^{j+1} + c_i^j - c_{i+1}^{j+1} - c_{i+1}^j\right)$$

$$- \frac{q_{i+1/2}^{j+1/2}\left(c_i^{j+1} + c_i^j\right) - q_{i-1/2}^{j+1/2}\left(c_{i-1}^{j+1} + c_{i-1}^j\right)}{2(\Delta z)} \qquad [5.5]$$

where the indices i and j, indicate sequences in space (z) and time (t), respectively, and $N_{i+1/2}^{j+1/2}$ is known as the numerical dispersion term and is defined as:

$$= \frac{\Delta z}{2} q_{i+1/2}^{j+1/2} \frac{V_i^{j+1/2} c_{i+1/2}^{j+1/2} - V_{i+1/2}^{j+1/2}\Delta t}{\Delta t}\left(\theta_i^{j+1} - \theta_i^j\right) \qquad [5.6]$$

and $\overline{V_i}$ and $\theta_{i+1/2}^{j+1/2}$ are defined by:

$$V_i = q_i^j/\theta_i^j \text{ and } \theta_{i+1/2}^{j+1/2} = \tfrac{1}{4}\left(\theta_i^{j+1} + \theta_i^j + \theta_{i+1}^{j+1} + \theta_{i+1}^j\right)$$

The term D is the hydrodynamic dispersion coefficient. The term N has the effect of an additional dispersion coefficient. This equation is implicit, due to the presence of the j+1 index on the right hand side of the equation, and can be solved by use of a tridiagonal matrix (Bresler et al. 1982).

If the moisture content and the flow rate are held constant for space and time, then hydrodynamic dispersion (D) and the correction term (N) will also remain constant and Eq. (5.4) becomes (Bresler 1973):

$$\frac{\partial c}{\partial t} = D^{*}\frac{\partial^2 c}{\partial z^2} - V^{*}\frac{\partial c}{\partial z} \qquad [5.7]$$

where D^{*} and V^{*} represent D/θ and V/θ (interstitial flow velocity) respectively. The finite difference form of this is:

$$c_i^{j+1} = c_i^j + \frac{\Delta t}{\Delta z}\left[\frac{D^{*}-N}{\Delta z}\left(c_{i-1}^j - 2c_i^j + c_{i+1}^j\right) - q^{*}\left(c_i^j - c_{i-1}^j\right)\right] \qquad [5.8]$$

which is essentially the form used by Leistra (1973) with the exception that in the place of θ Leistra (1973) used a volumetric pesticide capacity factor, ψ, which was defined as:

$$\psi = \theta + \rho_b K_{s/w} \qquad [5.9]$$

where ρ_b is the bulk density and $K_{s/w}$ is the soil/water distribution ratio of the pesticide representing its mobility. Because the solute used in this study, oxygen-18 behaves like normal water, oxygen-16, volumetric moisture content, θ, only is used as a denominator for D and q. For ease of calculation Eq. (5.8) is represented by the following (Leistra 1973; Miller 1983):

$$C_{i,j+1} = (r_v + r_d)C_{i-1,j} + (1 - r_v - r_d - r_d)C_{i,j} + (r_d)C_{i+1,j} \qquad [5.10]$$

where the indices i and j, indicate sequences in space, Δz, and for time, Δt, respectively; and $r_v = (v/\theta)(\Delta t/\Delta z)$ and $r_d = \Delta t((D-N)/\theta)/(\Delta z)^2$. Here r_{d_i} refers to the spreading from compartment i in the forward direction. Initial conditions define

C_{i,j_0} for all z . The concentration for the next time step, C_{i,j_1}, is determined from C_{i,j_0} at each depth increment and the process repeated for each time step. Thus the mass of solute transported is proportional to the volume of water moving between layers and the solute concentration in the water. This numerical solution is based upon backward differences rather than central differences. Correction for numerical dispersion is approximated by N (Bresler 1973):

$$N = 0.5\, q\, \Delta z \qquad [5.11]$$

with the condition that $N \leq D$. To solve equation [5.10] for stable conditions, it is essential that the following conditions hold (Leistra 1973):

$$r_v + 2r_d < 1 \qquad [5.12]$$

Equation [5.10] may be extended so that different moisture contents and dispersion coefficients are possible within the soil profile. The soil profile is divided into a series of compartments in which the moisture content and hydrodynamic dispersion are different from within other compartments. For such conditions the following equation is used (Leistra 1973):

$$C_{i,j+1} = (r_v + r_{d\,i-1})C_{i-1,j} + (1 - r_v - r_{d\,i-1} - r_{d\,i})C_{i,j} + (r_{d\,i})C_{i+1,j} \qquad [5.13]$$

where for r_v and r_d, the moisture content is that of the current layer, θ_i. Flux may may be of different values between compartments only if the excess water resulting is removed at the boundary between the two different fluxes. Such a condition of water removal allowing a different flux below and above is that of moisture withdrawal mid-profile due to roots.

5.3. CONDITIONS OF SENSITIVITY ANALYSIS

5.3.1 Types of Flow Conditions Simulated.

Four types of conditions were simulated and evaluated with a sensitivity analysis:

1. **Model 1**; no seasons; constant flow in one direction only, downward from surface to water table at 400 cm.
2. **Model 2**; no seasons; constant flow to bottom of root zone at 60 cm, downward from surface and upward from water table at 400 cm.
3. **Model 3**; seasonal flow, removal from soil surface; downward constant flow from surface to water table during one season and upward constant flow from water table to soil surface during other season.
4. **Model 4**; seasonal flow, removal from bottom of root zone at 60 cm; downward constant flow from surface to water table during one season and upward constant flow from water table to root zone during other season.

The four conditions represent the following types of possible 'situations' respectively; (1) constant downward leaching of solutes, (2) constant plant growth with moisture supplied from surface and from the water table, (3) seasonal flow with drainage during one season and the capillary fringe supplying water to an evaporating soil surface during the next, and (4) seasonal flow with drainage during one season and the capillary fringe supplying water to a transpiring root zone or a freezing front during the following season. The main criterion for this simulation is that the water input from the surface has a different isotope concentration than the groundwater.

5.3.2 Initial and Boundary Conditions

The physical system simulated occupies the region $0 \leq z \leq L$, where the water table is at L+1. Solute is allowed to flow either from a large reservoir at $z = 0$ or $z = L+1$ into a column at a constant rate of $q\theta^{-1}Co$, where q is the water velocity (cm/d), θ the volumetric moisture content (m^3/m^3) and Co the concentration of the influent

expressed in terms of relative oxygen-18 concentration (‰). The solute is allowed to diffuse and disperse in the region $0 < z \leq L$. The upper boundary and lower boundaries, $z = 0$ and $z = L+1$, are represented by infinitely large pools of water of constant solute concentrations of, C_s and C_w, respectively, moving into the soil system with an initial concentration of C_p. Calculations begin at $0 < z$ or at $z = L$ (for upward flow) where the $C_{i-1,j}$ term in Eq. [5.13] is replaced by C_s or C_w.

The following initial and boundary conditions for the various models were used:

1. Model 1; no seasons; constant downward flow:

C_s = -15 ‰	at $z = 0$	for $0 \leq t$
C_p = -16 ‰	at $0 < z \leq L$	for $t = 0$
C_w = -17 ‰	at $z = L + 1$	for $0 \leq t$
q_s = downward flux (+)	at $0 < z \leq L$	for $0 \leq t$

2. Model 2; no seasons; constant flow to root zone:

C_s = -15 ‰	at $z = 0$	for $0 \leq t$
C_p = -17 ‰	at $0 < z \leq L$	for $t = 0$
C_w = -19 ‰	at $z = L + 1$	for $0 \leq t$
q_s = downward flux (+)	at $0 < z \leq R$	for $0 \leq t$
q_w = upward flux (-)	at $R \leq z \leq L$	for $0 \leq t$

3. Model 3; seasonally alternating flow; removal from soil surface:

C_s = -15 ‰	at $z = 0$	for $0 \leq tn$
C_p = -17 ‰	at $0 < z \leq L$	for $t = 0$
C_w = -19 ‰	at $z = L + 1$	for $0 \leq tn$
q_s = downward flux (+)	at $0 < z \leq L$	for $0 \leq t_{2n+1}$ (odd seasons)
q_w = upward flux (-)	at $0 < z \leq L$	for $0 \leq t_{2n}$ (even seasons)

4. Model 4; seasonally alternating flow; removal from root zone (R):

C_s = -15 ‰	at $z = 0$	for $0 \leq tn$
C_p = -17 ‰	at $0 < z \leq L$	for $t = 0$
C_w = -19 ‰	at $z = L + 1$	for $0 \leq tn$
q_s = downward flux (+)	at $0 < z \leq L$	for $0 \leq t_{2n+1}$ (odd seasons)
$q_{w1} = 0$	at $0 < z < R$	for $0 \leq t_{2n}$ (even seasons)
q_{w2} = upward flux (-)	at $R \leq z \leq L$	for $0 \leq t_{2n}$ (even seasons)

For downward flow, the solute reaching the water table (z = L+1) is considered mixed with an infinitely large reservoir of concentration C_w. A winter 'no-flow' soil layer with removal at the bottom of the root zone would essentially be the same as allowing downward flow from the surface and removal of this water by transpiration at the root zone (R) along with the upward flowing groundwater.

For models 3 and 4, the following test is performed at the end of each year (i.e. at the end of the upward flow season, winter, or at the beginning of the downward flow season, summer):

$$\sum_{i=1}^{L} \left(|C_{i,s} - C_{i,s-1}| + |C_{i,w} - C_{i,w-1}| \right) \leq 0.01 \qquad [5.14]$$

where the s and w subscripts represent the solute concentration for the last time step at the end of the summer and winter seasons respectively. Equation [5.14] is used to check whether stability has been achieved. Calculations for each simulated year are repeated until tn, when the total difference is equal to or less than 0.01 or some other selected value. Conditions of equilibrium for model 2 are similar to Eq. [5.14] with the exception that the winter terms $C_{i,w}$ are removed. As model 1 is constant flow in one direction only, equilibrium would be achieved when the surface solution, C_s, finally replaced the initial solution for the entire profile. Thus for the purposes of the sensitivity analysis, model 1 is run for a specified period of time.

5.3.3 'Standard Conditions' For Sensitivity Analysis.

The simulated physical system has a profile depth of 402 cm (for a Δz of 6 cm), with a water table located at the bottom of the profile. Simulated flow rates are very low, 0.01 to 0.3 cm/day, which might occur under conditions of evapotranspiration, flow to a freezing front, or drainage during unsaturated conditions. Moisture contents

represent those of a clay loam system, ranging from 0.20 for dry conditions to 0.40 for saturated conditions. A soil temperature of 5 °C and a porosity of 0.40 were used for calculation of the diffusion coefficient. The standard values for both the one-directional and alternating direction models are listed in Table 5.1.

The input parameters used for the sensitivity analysis of Eq [5.13] in this section are: the depth increment, time step, moisture flux, apparent dispersion coefficient, moisture content, and length of run or relative length of seasons (for alternating flow). One input variable at a time is evaluated by varying the selected variable and holding the others constant.

Table 5.1. Standard values for model input parameters used in the sensitivity analysis.

Property	Value		
	Model 1	Models 2‡, 3, and 4	
		Summer	Winter
Flux; q (cm/d)	0.10	0.10	-0.10
Dispersion†; D* (cm^2/d)	0.74	0.74	0.74
Diffusion; D_s^{dif} (cm^2/d)	0.12	0.12	0.12
Moisture content; θ (cm^3/cm^3)	0.30	0.30	0.30
Length of seasons; Ts, Tw (d)	-	183*	183*
Length of run; T (days)	366	Until conditions for Eq. [5.14] are met	
Time step; Δt (d)	2	2	2
Depth interval; Δz (cm)	6	6	6
Input concentrations; C (‰)	-15	-15	-19
Concentration at t=o	-17	-17	-17

‡ For model 2 the summer conditions represent flow from the soil surface to the root zone and winter conditions flow from the water table to the root zone, with the exception of season length.

† Dispersion coefficient (D*) shown above is that calculated for a flux of 0.10 cm/d by Eq. [5.3] and corrected for numerical dispersion by Eq. [5.11].

* For alternating flow with moisture removal in the root zone the length of the seasons change to 200 days for summer and 166 days for winter.

For the sensitivity analysis, the variables describing the physical system (q, D^*, θ, Tw, Ts, and T) are varied by ± 25% of the standard values and the output oxygen-18 concentrations are compared against the standard run. There are two attributes of the output that should be measured; change in solute concentration and change in the shape of the distribution of solute concentration in the profile. The first will be evaluated by taking the maximum difference between that of the property varied and standard conditions, while the second, profile shape, is evaluated by comparing the depth of occurrence of the solute inflection point (i.e. where the solute concentration of -16‰ for model 1 or -17 ‰ for the other models occurs) and the immediate slope of the solute profile about the inflection point. Graphs will also be presented for visual analysis.

For moisture content, ± 25% of the standard value results in a greater range than that normally encountered for field measurements (Warrick and Nielsen 1980; Hawley et al. 1983). As the variables, q and D^*, can easily vary by more than 100% (Nielsen et al. 1973; Warrick and Nielsen 1980), they are further evaluated by varying their values from 0.2 to 2.0 times their standard values.

5.4. SENSITIVITY ANALYSIS; RESULTS

5.4.1 Diffusion and Dispersion Coefficients

The diffusion coefficient in Eq. [5.2], only exceeded the uncorrected dispersion in Eq. [5.3] for fluxes below 0.03 cm/d, dependent upon the moisture content. The diffusion coefficient decreases as the moisture content decreases for a fixed porosity (Fig. 5.1).

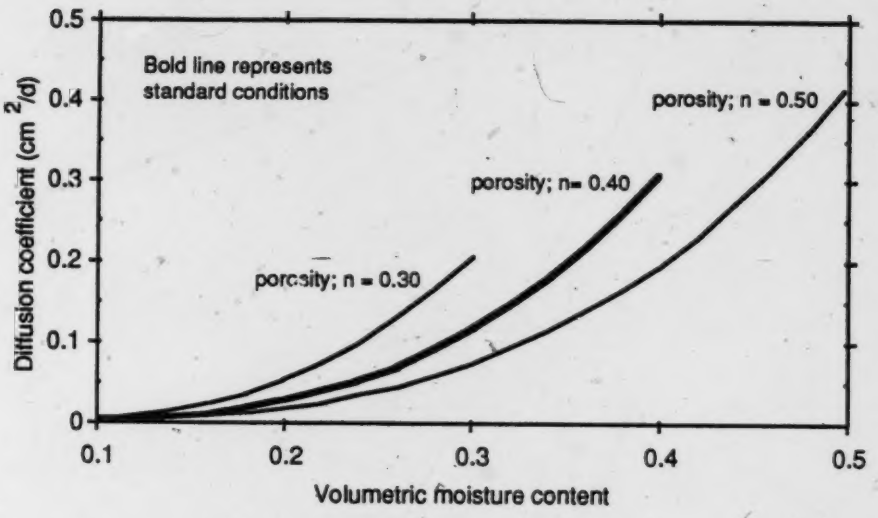

Fig. 5.1 Effect of moisture content and porosity upon diffusion (Equation 5.2).

The apparent hydrodynamic dispersion coefficient as a function of flux (q), uncorrected and corrected for numerical dispersion and for different depth intervals (Δz), is shown in Fig 5.2. Correcting hydrodynamic dispersion for numerical dispersion has the effect of reducing dispersion. The larger the depth interval, the smaller the corrected dispersion (D*). For specific depth intervals, the D* increases and then decreases to zero as flux (q) increases. For the conditions outlined in this chapter, the corrected dispersion used in Eqs. [5.10, 5.13] generally ranges between 0.02 and 1 cm^2/d.

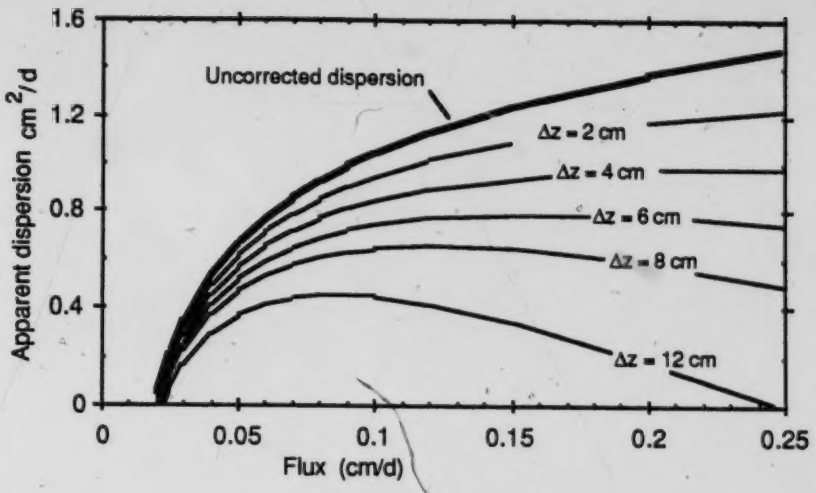

Fig. 5.2 Effect of depth interval (Δz) and flux (q) upon corrected dispersion (Equations 5.3, 5.6).

5.4.2 Model 1: constant flow downward.

The successful application of any finite difference model is dependent first upon the correct selection of the depth and time grid. Computer time can be made much shorter with a larger grid size, however, if the grid size is too large then the accuracy of the numerical solution decreases (Wang and Anderson 1979; Hillel 1980). Thus a check on grid size should be the first evaluation of any numerical model. For a dispersion model, the grid analysis also provides a check on the error term for numerical dispersion. The results of such an analyses are compared in Table 5.2 to those of the standard conditions (Table 5.1).

Table 5.2. Results of varying depth interval and time step in influencing simulated oxygen-18 concentrations.

Conditions Δz (cm)	Δt (d)	Max. abs difference ‰	Depth and slope for -16 ‰ cm	Δ‰/Δz
2	0.5	0.005	132	-0.0167
3	1.0	0.003	132	-0.0167
6	1.0	0.005	132	-0.0167
6	3.0	0.005	132	-0.0167
6	2.0	norm	132	-0.0167
12	8.0	0.007	132	-0.0167

† Standard conditions are listed in Table 5.1. The maximum absolute difference is that which occurs between the simulation for the property varied and the standard conditions at any individual depth. Depth and slope represent where the concentration value of -16‰ is first encountered. Depth intervals for the different depth intervals were interpolated to match that of 6 cm.

For the grid size analyzed, a depth interval from 2 to 6 cm, and a time interval from 1 to 3 days for a Δz of 6 cm, there was no difference in output concentrations (Table 5.2). Any of the depth and time steps within the range considered in Table 5.2 are permissible for constant flow in one direction. Leistra (1973) found the correction factor for numerical dispersion to result in more exact solutions for small time steps, however, for the model considered in this paper, changing the timestep resulted in virtually no change in solution concentration.

Varying the flux by ± 25% of 0.10 cm/d (0.075 cm/d and 0.125 cm/d) resulted in a maximum difference of about 0.5‰, whereas flux variations close to 100% (0.02 cm/d and 0.20 cm/d) resulted in differences close to the maximum possible, 2‰ (Table 5.3, Fig. 5.3). The greater the flux, the greater the penetration and the steeper the slope that occurs for a concentration of -16‰. The steeper slope for increasing flux is indicative of the greater dispersion that takes place.

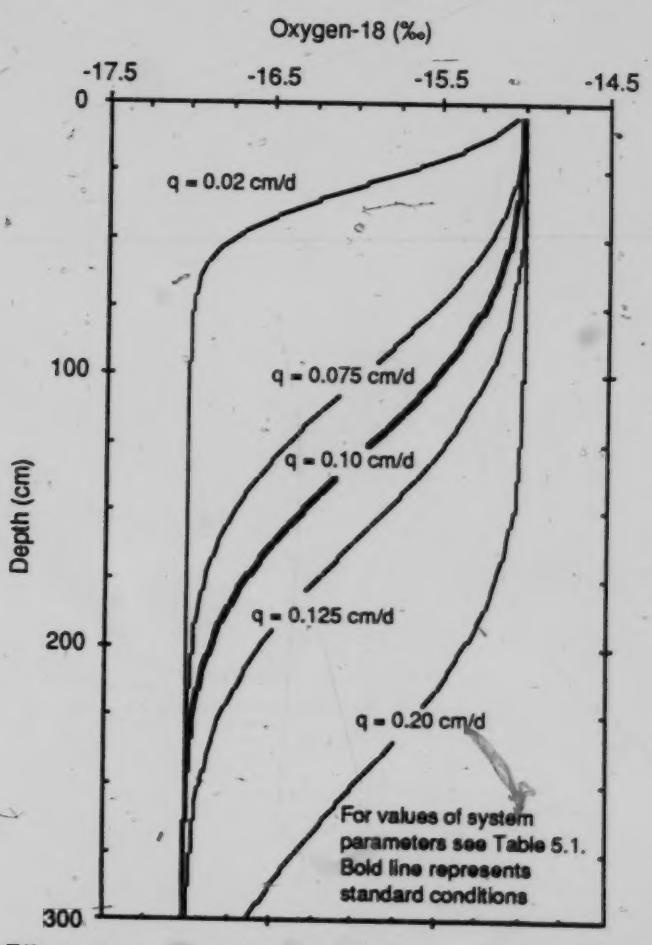

Fig. 5.3. Effect of varying moisture flux upon simulation of oxygen-18 distribution for conditions of downward flow.

Varying the dispersion by ± 25% of 0.74 cm^2/d results in the least change of the output concentrations for any of the properties varied (Table 5.3). Changing the dispersion affects the spreading of the front and not the depth of penetration, as evidenced by the steeper slopes with increasing dispersion, (Table 5.3, Fig. 5.4). Due to the effects of numerical dispersion, a coefficient of zero dispersion does not result in a 'piston-like' flow of solutes (Fig. 5.4). This effect of numerical dispersion poses a problem only if conditions of zero dispersion are modelled.

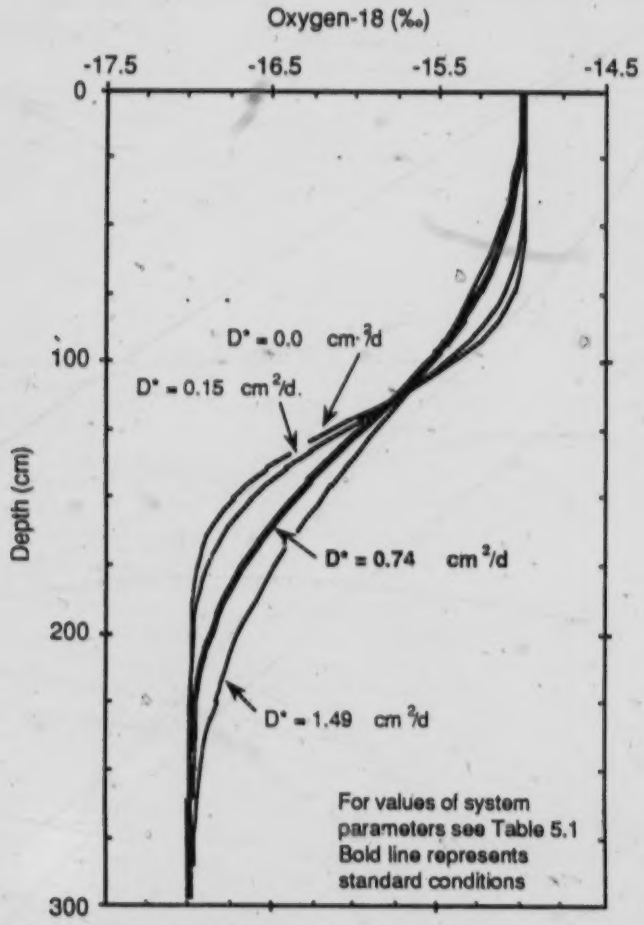

Fig. 5.4. Effect of varying apparent hydrologic dispersion upon simulation of oxygen-18 distribution for conditions of downward flow.

Table 5.3. Results of varying depth interval and time step in influencing concentrations and dispersion of oxygen-18 as compared against standard conditions for a simulated soil profile

Conditions	Max. abs difference	Depth and slope for -16 ‰ cm	Δ‰/Δz
Standard conditions	-	132	-0.0167
Flux; q (cm/d)			
0.010 (T=4000 d)	0.21	144	-0.0149
0.020	1.80	36	-0.0458
0.075	0.52	108	-0.0181
0.125	0.48	162	-0.0158
0.200	1.52	252	-0.0147
Dispersion; D* (cm^2/d)			
0.000	0.38	126	-0.0308
0.149	0.27	126	-0.0253
0.558	0.06	132	-0.0184
0.931	0.05	134	-0.0155
1.489	0.18	138	-0.0132
Moisture content; θ (cm^3/cm^3)			
0.225	0.63	174	-0.0145
0.375	0.44	108	-0.0188
Length of run; T (d)			
276	0.54	102	-0.0196
456	0.49	168	-0.0148

† Standard conditions are listed in Table 5.1. The maximum absolute difference is that which occurs between the simulation for the property varied and the standard conditions at any individual depth. Depth and slope represent where the concentration value of -16 is first encountered. Depth intervals for the different depth intervals were interpolated to 6 cm.

Varying the moisture content essentially affects both the the depth of solute penetration and the spreading of the solute profile as evidenced by the steeper slopes (Table 5.3, Fig. 5.5). The moisture content affects the pore water velocity and the dispersion term, Eqs [5.10, 5.13]. The lower the moisture, the greater the pore water velocity (q/θ) and the greater the dispersion (D/θ). Varying the moisture content by ± 25% had a similar effect to that of varying the flux by ± 25% (Table 5.3, Figs 5.3 and 5.5).

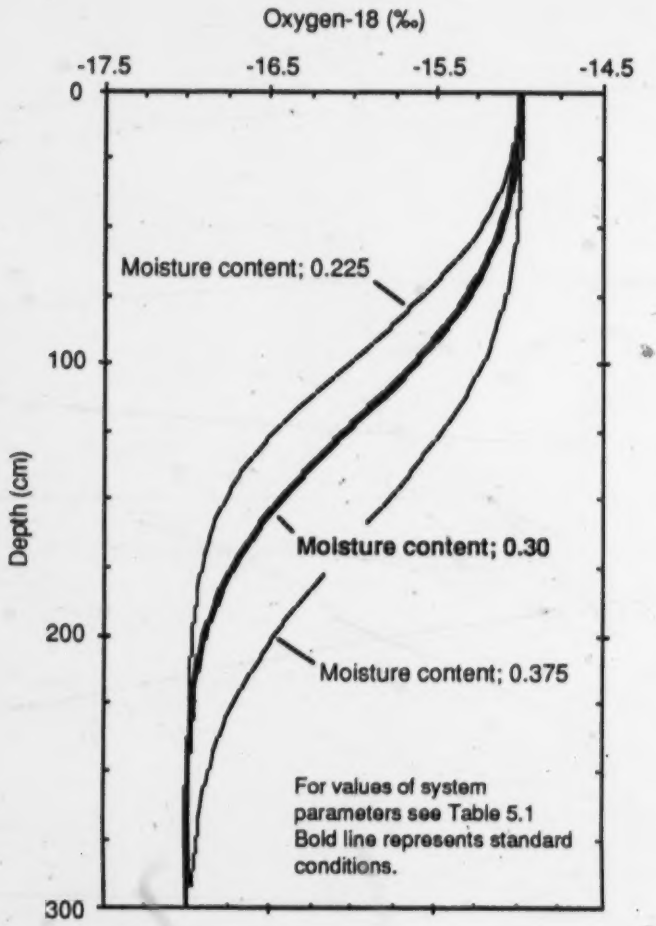

Fig. 5.5. Effect of varying water content upon simulation of oxygen-18 distribution for conditions of downward flow.

As with moisture content and with flux, varying the length of run by ± 25% had a similar effect to the oxygen-18 concentrations as compared to the standard (Table 5.3, Fig. 5.6); the longer the run the greater the penetration and the more the spreading, resulting in a steeper slope.

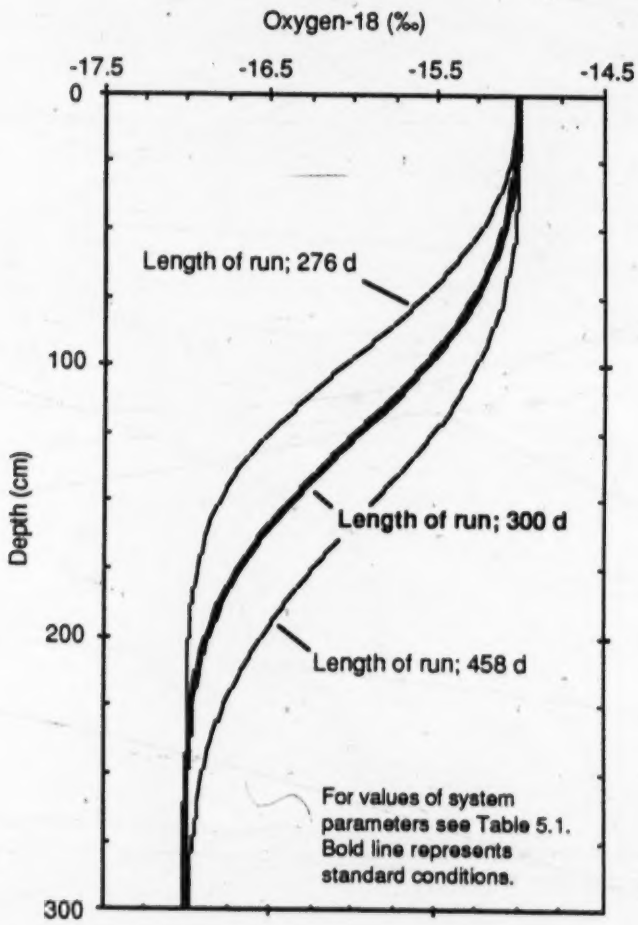

Fig. 5.6. Effect of varying length of time period upon simulation of oxygen-18 distribution for conditions of downward flow.

In summary, for constant flow in one-direction, the dispersion coefficient has the least effect upon the output concentrations as compared to the other input parameters. For model applications this is fortunate as dispersion is difficult to measure and is highly variable. As moisture content may be accurately measured and generally has a low field variability, it is thus an easily controlled parameter. Flux represents the hardest input parameter to control for several reasons: it has a large effect upon the output; it is difficult to accurately measure; and it has a large field variability.

5.4.3 Model 2: Flow towards a root zone.

Varying the input parameters by ± 25% of the standard values results in the same degree of differences as for Model 1 (compare Tables 5.3 and 5.4). The main difference is that of upward flow from the water table with removal from the root zone at 60 cm. Generally for all variations of flux, dispersion, and moisture content, profile concentrations in model 2 will eventually obtain the same shape. This is approximated by the 1500 day run for an upward flow rate of -0.10 cm/d as illustrated in Fig. 5.7. For very low flow rates of 0.01 cm/d, this same profile shape takes longer than 100 years to achieve. Dispersion and diffusion allows for some amount of spreading on both sides of the zone of moisture removal at 60 cm (Figs. 5.7 and 5.8). This is especially evident in Fig. 5.8 where for zero dispersion there is no spreading across the 60 cm depth where moisture is withdrawn.

Table 5.4. Sensitivity analysis of input parameters for Model 2.

Conditions	Max. abs difference	Depth and slope for -18 ‰ cm	Δ‰/Δz
Standard conditions	-	264	-0.016
Flux; q (cm/d)			
0.020	1.84	378	-0.046
0.075	0.54	300	-0.018
0.125	0.51	234	-0.015
0.200	1.57	132	-0.014
Dispersion; D* (cm^2/d)			
0.000	0.39	276	-0.030
0.149	0.27	278	-0.024
0.558	0.07	264	-0.018
1.489	0.19	258	-0.012
Moisture content; θ (cm^3/cm^3)			
0.225	0.66	222	-0.014
0.375	0.46	294	-0.018

† Standard conditions are listed in Table 5.1. The maximum absolute difference is that which occurs for the property varied and the standard conditions at any individual depth.

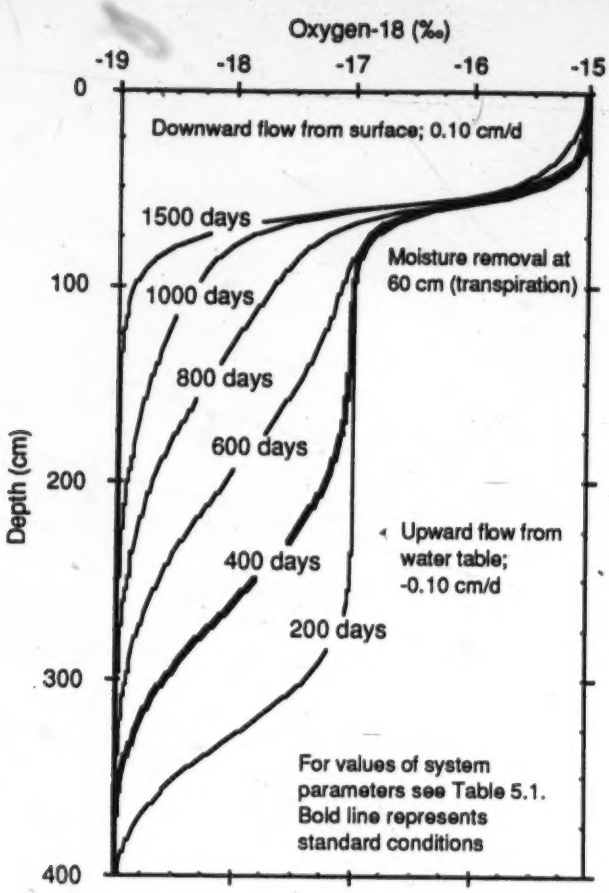

Fig. 5.7. Effect of varying length of time period upon simulation of oxygen-18 distribution for conditions of upward flow to a root zone.

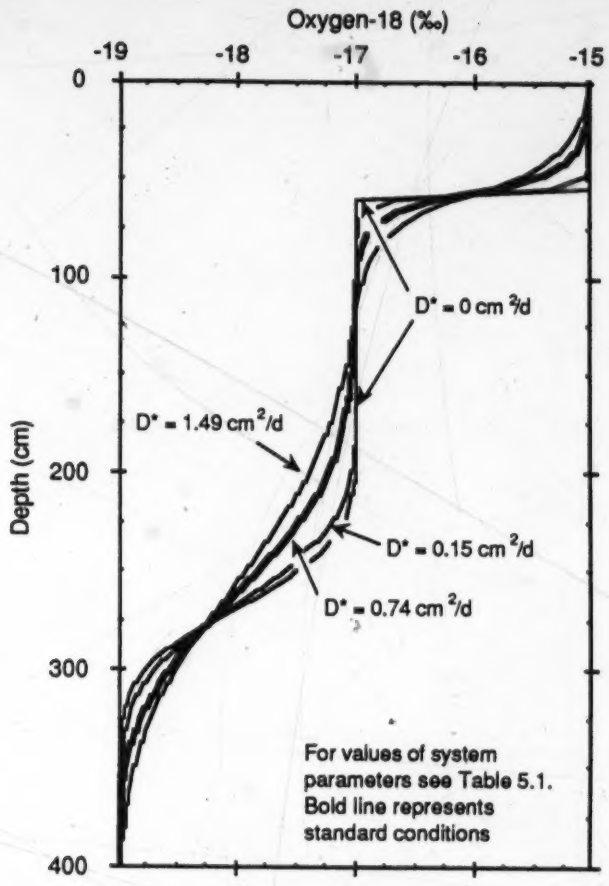

Fig. 5.8. Effect of varying dispersion upon simulation of oxygen-18 distribution for conditions of upward flow to a root zone.

5.4.4 Model 3: Seasonal flow, moisture removal from the surface.

In this model moisture removal during conditions of upward flow (winter) was from the soil surface. Sensitivity analysis of the model was performed for changes in grid size, flux, moisture content, dispersion, and relative season length. The model was run until equation 5.14 was satisfied. For the cases considered in this section, simulated time ranged between 20 and approximately 150 years, with low flow rates taking longer to equilibrate.

The major difference in the output of this model over that of the previous ones is that there are two profiles; one for winter and one for summer (Figs. 5.9 to 5.11). This is due to the alternating flow directions. The concentrations plotted in each figure occur at the end of each season. As upward flow occurs during the winter season, the oxygen-18 values are larger for specific depths than those for summer when downward flow occurs. The winter profile has a convex shape while the summer is concave.

With the exception of relative season length and different seasonal fluxes, the sensitivity analysis indicates that the input parameters have less of an effect upon the oxygen-18 concentration and distribution than one-directional flow for model 1 had (compare Table 5.3 with Table 5.5, and Fig. 5.3 with Fig. 5.9). This is due to the fact that the models were run until equilibrium had been reached and thus dispersion was allowed to reach its maximum effect as evidenced by the near linear slope for the profile concentration distribution (Fig. 5.9).

Table 5.5. Results of varying model input parameters upon concentrations of oxygen-18 as compared against standard conditions for the summer season of the alternating flow model, with surface removal.

Conditions ‰	Max. abs difference	Depth and slope for -17‰	
		cm	Δ‰/Δz
Standard conditions	-	240	-0.0103
Depth interval; Δz (cm)			
2 (Δt; 0.5 d)	0.06	234	-0.0106
3 (Δt; 1.0 d)	0.05	234	-0.0106
12 (Δt; 3.0 d)	0.01	240	-0.0101
Time step for Δz = 6 cm; Δt (d)			
1	<0.01	240	-0.0103
3	<0.01	240	-0.0103
Absolute flux; q (cm/d)			
±0.020	0.46	216	-0.0029
±0.075	0.12	228	-0.0100
±0.125	0.13	246	-0.0107
±0.200	0.53	270	-0.0120
Winter flux; q (cm/d)			
-0.075	1.93	278	-0.0279
-0.125	1.97	90	-0.0271
Moisture content; θ (cm^3/cm^3)			
0.225	0.17	246	-0.0108
0.375	0.10	234	-0.0101
Dispersion; D* (cm^2/d)			
0.000	0.05	240	-0.0100
0.149	0.05	240	-0.0102
0.558	0.01	240	-0.0103
0.931	0.01	240	-0.0103
Relative length of seasons; summer/winter (d)			
156/210	2.06	96	-0.0282
210/156	2.02	384	-0.0368

† Standard conditions are listed in Table 5.1. The maximum absolute difference is that which occurs between the simulation for the property varied and the standard conditions at any individual depth. Depth and slope represent where the concentration value of -16 is first encountered. Depth intervals for the different depth intervals were interpolated to match that of 6 cm. With the exception of the depth of occurrence of the -17‰ value all winter absolute differences and slopes were the same as the summer values.

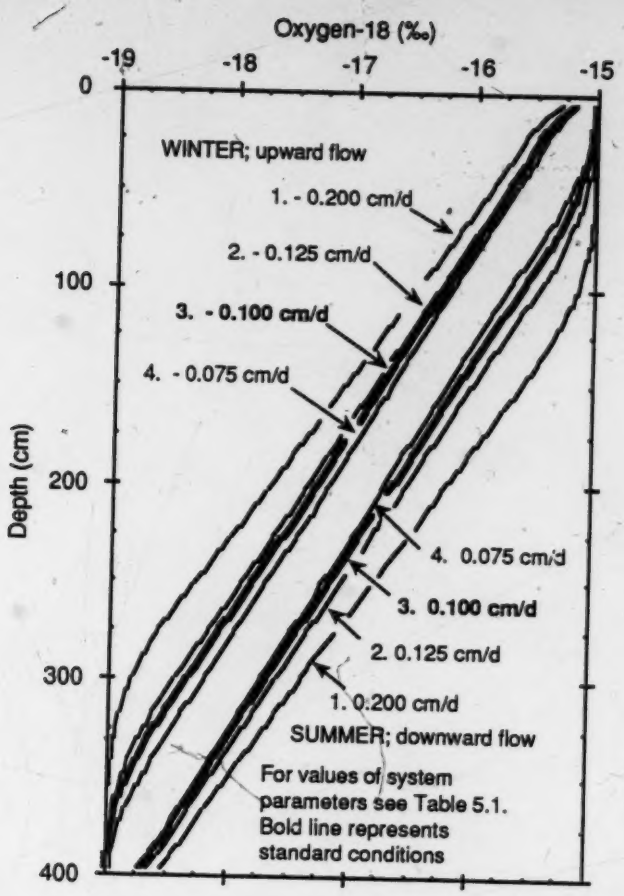

Fig. 5.9. Effect of varying absolute moisture flux upon simulation of oxygen-18 distribution for conditions of alternating flow with removal from the surface.

Variation in grid size and in the dispersion coefficient had virtually no effect upon the oxygen-18 concentrations (Table 5.5). Increasing the absolute moisture fluxes, while keeping them the same between seasons had the effect of increasing the difference between the seasons (Fig. 5.9), with the slope slightly decreasing and depth of occurrence of -17 ‰ increasing as the absolute flux increased (Table 5.5). Varying the absolute flux for alternating flow has less of an effect upon the profile oxygen-18 concentrations, than the same fluxes had for Model 1.

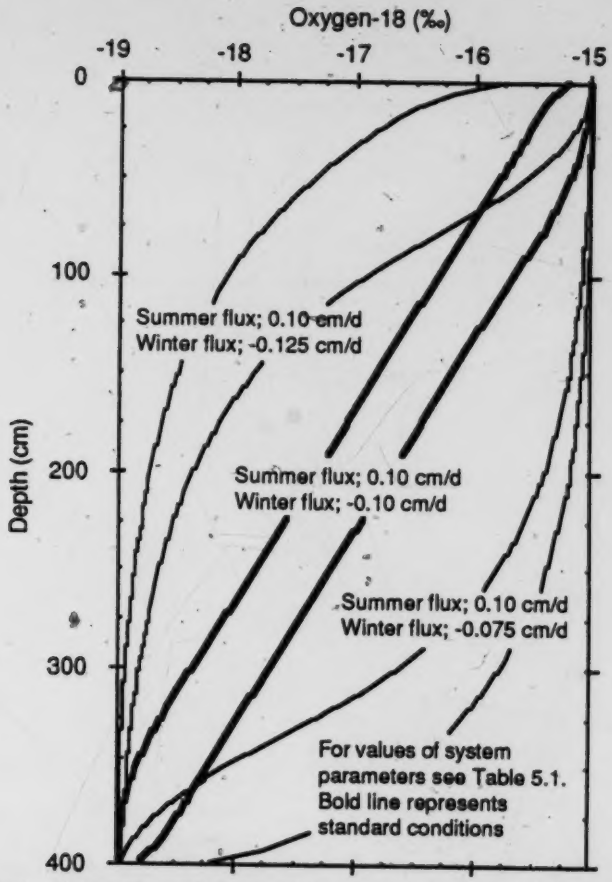

Fig. 5.10. Effect of varying relative moisture flux upon simulation of oxygen-18 distribution for conditions of alternating flow with removal from the surface.

Varying the winter moisture flux while maintaining summer flux at a constant rate of 0.10 cm/d (i.e. changing the relative moisture flux) results in large differences from that of standard conditions for a small difference between seasonal flux (Fig. 5.10, Table 5.5). These large differences are due to the simulated long time period. When the upward flux of the winter season is greater than that of the summer, then the profile will have lower concentrations than standard conditions as upward flow dominates (Fig. 5.10). Likewise if summer has a higher flux rate, then the profile will have higher concentrations. The greater the difference in seasonal fluxes the greater the convexity or concavity of the resulting profile concentrations.

Varying the length of the seasons, while holding the seasonal fluxes the same, has similar effects; a longer summer will result in a profile of higher concentrations than that of the standard conditions (Fig. 5.11, Table 5.5).

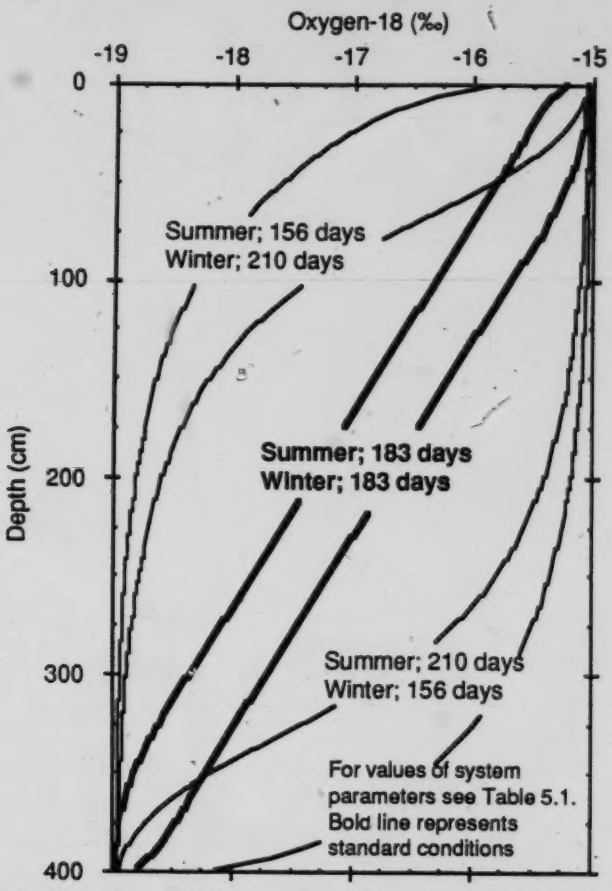

Fig. 5.11. Effect of varying relative season length upon simulation of oxygen-18 distribution for conditions of alternating flow with removal from the surface.

Varying the moisture content has an effect similar to that of varying the absolute flux; that of increasing the difference between the summer and winter concentrations but not varying the slope to any great extent (Table 5.5). The range of moisture contents considered (± 25%) resulted in very little difference from that of the standard conditions.

5.4.5 Model 4: Seasonal flow, moisture removal from the root zone.

Removing soil moisture from the 60 cm depth, rather than the surface, has the effect of creating a 'false' surface at 60 cm depth (Figs. 5.12 and 5.13). Other than this, there are no other major differences from the alternating flow model with moisture removed from the surface. In both models 3 and 4 decreasing the overall moisture flux results in decreasing the difference between seasonal concentrations, but the general profile shape is maintained (Figs 5.12 and 5.13).

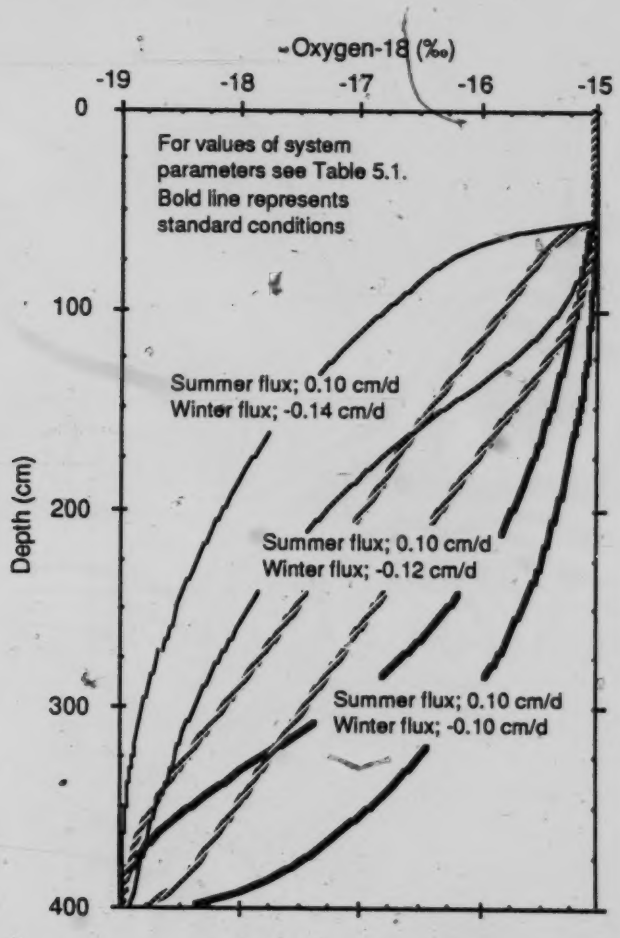

Fig. 5.12. Effect of low moisture fluxes (>0.10 cm/d) upon simulation of oxygen-18 distribution for conditions of alternating flow with removal from the root zone.

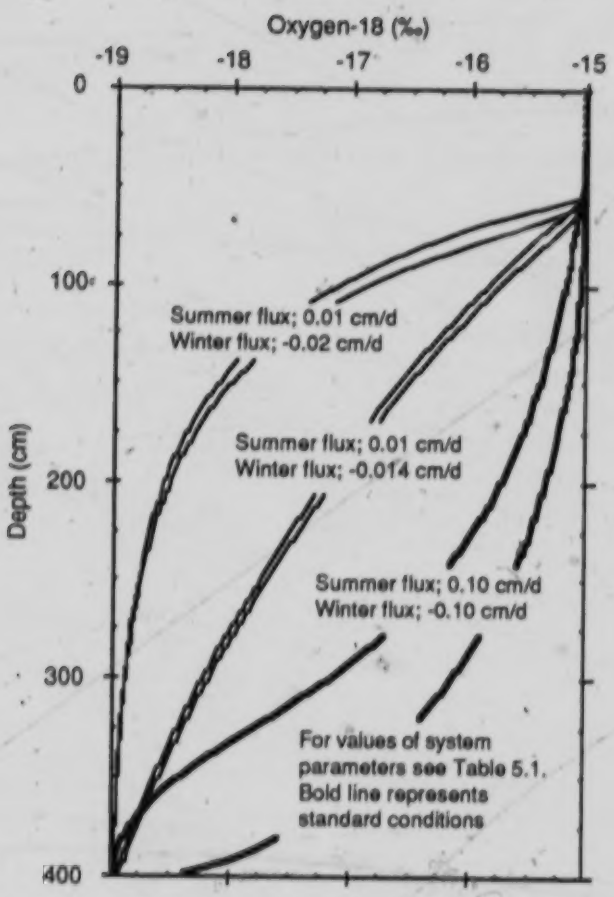

Fig. 5.13. Effect of very low moisture fluxes (<0.02 cm/d) upon simulation of oxygen-18 distribution for conditions of alternating flow with removal from the root zone.

5.5 SUMMARY AND CONCLUSIONS

The purpose of the paper was to develop and evaluate a dispersion model that simulates seasonal water movements that may be typical for the Canadian Prairies. Four types of flow conditions were simulated for a convective flow-dispersion model for non-reactive solutes. The different flow conditions were evaluated with a sensitivity analysis. The flow conditions were: **Model 1**, constant flow downwards; **Model 2**, constant flow down from the surface and up from a water table to the root zone; **Model 3**, seasonal flow alternating in direction with removal from the surface; and **Model 4**, seasonal flow alternating in direction with removal from the root zone. The flow conditions are low flow rates resulting from unsaturated drainage, capillary rise to a surface evaporating front or to an actively transpiring root zone, or moisture flux to a freezing front.

The main difference among the four models was that Models 3 and 4 had seasonally alternating flow, resulting in two profile concentrations; one for the downward flow season, and one for the upward flow season. The four conditions also have a similar partitioning on the basis of sensitivity analysis, with Models 1 and 2 forming one group and 3 and 4 the other. This is primarily due to the fact that Models 1 and 2 were run for a limited time, usually only 400 days, while Models 3 and 4 were run until a steady state equilibrium between the seasons had been reached, usually about 30 to 200 years. Varying the dispersion had the least effect for all the models, while varying the flux and the moisture content had large effects for the first two models. All input parameters were varied by ± 25% of the standard values; with the following results:

Model 1; moisture content and flux had the greatest, and almost equivalent, effect upon output concentrations. As both the flux and dispersion terms in the model are divided by moisture content, this is not surprising. For field measurements moisture content is an simply and relatively accurately measured property and thus this parameter is easily controlled. Flux and dispersion are difficult to measure and also have high field variations. Fortunately dispersion has the smallest effect upon output concentrations. If the model is to be used to predict profile concentrations, then the flux term can be the largest source of error resulting in a possible difference of more than 50% in concentration for a flux value 100% different from reality. However, if the concentrations and moisture contents are known for a profile under downward flow conditions, then it should be possible to estimate the flux to within ±25% of the actual value.

Model 2; Constant flow towards a transpiring root zone results in a sharp change in profile concentrations at the depth of moisture removal in the root zone. Dispersion results in some amount of spreading on either side of the root zone. Sensitivity analysis indicated that varying the input parameters had the same degree of effect upon the profile concentrations as in Model 1.

Models 3 and 4; The basic difference in concentration profiles between these two models is that the water is removed from the surface for Model 3, whereas for Model 4 the water is removed from the root zone at 60 cm. Concentration profiles are different for the upward flow season as opposed to the downward flow season. This difference between seasons decreases as the overall flux decreases. Steady state conditions take between 30 and 200 years to achieve, with the time increasing for the smaller fluxes. Varying the dispersion coefficient, the absolute moisture flux, and the moisture content

has much less of an effect upon the output for the same amount of variations for Models 1 and 2. However, varying the relative moisture fluxes between seasons or the relative lengths of season result in large differences in profile shape and concentrations of output. If Model 4 is to be used to aid in the prediction of moisture flux, given that moisture content, solute concentrations and relative season lengths are known, it may prove difficult to determine the absolute moisture fluxes to any greater accuracy than an order of magnitude; however, the relative moisture fluxes (i.e. that one is greater than the other) may be determined to within 50% of the other.

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6. SIMULATION OF SEASONALLY VARYING ONE-DIMENSIONAL SOLUTE TRANSPORT: II. APPLICATION OF FIELD DATA.

6.1 INTRODUCTION

The distinct seasonality of the prairies, with its long winter, and warm dry summer temperatures, is reflected in numerous ways in the soil and shallow groundwater zones.

During much of winter, the ground on the prairie may be frozen to a depth of two meters (Maclean 1974) and movement of water from the water table to the freezing front is not uncommon (Willis et al. 1964; Benz et al. 1968; Cary et al. 1979; Gray et al. 1983). Field studies by Willis et al. (1964) and Benz et al. (1968) found that a drop in water table level occurred during winter along with increases in the soil moisture content. Cary et al. (1979) calculated that moisture flow to the freezing front for field conditions varied between less than 0.05 and 0.5 mm/day. Gray et al. (1983) measured overwinter moisture increases in the top meter of 47.8 mm (0.46 mm/d) and 95.7 mm (1.10 mm/d) for a silty clay soil with a water table below 4 m and for a fine sandy loam soil with a water table at 2.8 m respectively.

Meltwater does not penetrate the soil deeply until the ground has thawed, which does not occur until most of the snow cover has melted (Granger et al. 1977; Gray et al. 1983). Thus much of the accumulated snow can occur as runoff. Runoff from stubble and tilled fields commonly varies between 50% and 90% of the snow water equivalent (Willis et al. 1961; Willis et al. 1969; Patterson 1985). Granger et al. (1984) found that snowmelt in severely or heavily cracked fields, common in heavy-textured clays subjected to continuous cropping, tended to enter the soil through the cracks. Very little runoff was observed and also very little wetting of the soil between the cracks was

measured. In uncracked frozen soils due to the poor infiltration rates it appears that very little of the meltwater makes its way directly into the groundwater system underlying agricultural fields, but rather might recharge through small depressions (Meyboom 1966).

Upward water flow to the root zone can also take place due to the presence of dry conditions during the summer. Gardner (1958) theorized that in the presence of an actively transpiring crop, the potential at the bottom of the root zone may be taken as the upper boundary for a shallow water table. Even in the absence of a shallow water table upward flow rates to the root zone of 0.4 cm/day and 0.2 cm/day have been reported by van Bavel et al. (1968) and Stone et al. (1973).

Water flow between the root zone and a shallow groundwater table (≤ 5 m) may alternate seasonally in direction of flow. This has been documented for prairie sloughs where springmelt water, accumulating in depressions, results in downward flow during spring and and upward flow during the summer as the slough dries and the surrounding plant community extracts water from the soil (Meyboom 1966). Accumulation of water in a soil zone due to upward flow from freezing may drain back down upon thaw (Schneider 1961). This phenomenon is thought to cause overwinter salt accumulation in the surface horizons (van Schaik and Rapp 1970; Cary and Mayland 1972; Landsberg 1981) which may be flushed back down due to thaw and ensuing precipitation (Truss 1959; Cary et al. 1979; Landsberg 1981).

The quantification of flow rates and directions with normal techniques, such as soil moisture and depth to water table measurements is difficult, as these techniques measure changes in storage from which direction of flow is inferred. The use of tracers

in the form of existing salts in solution (Cary et al. 1979), nitrates (Cameron et al. 1979), chlorides (Schulin et al. 1987) radioactive isotopes (Cameron and Wild 1982; Miller 1983) or environmental isotopes (Saxena 1984; Bengtsson et al. 1987) not only enables estimates of flow rates and directions but also mixing of different waters. For the study of seasonal water flows the environmental isotopes, deuterium and oxygen-18, have the advantage over the other tracers of naturally varying in seasonal abundance. Thus summer and winter precipitation may be thus distinguished from each other. Knowledge of the amount and seasonal nature of water flow between the soil and groundwater systems is vital if nutrient losses through leaching, contaminant movement, salt fluxes, and groundwater contribution to plant growth are to be measured and understood.

The objectives of this paper are to quantify seasonal soil water flow directions and rates using stable isotopes with a numerical model. Measured soil moisture and oxygen-18 data obtained from a two year field study of overwinter changes in soil moisture resulting from snowmelt and groundwater contributions, are analyzed with a convection-dispersion numerical model. Flow rates, directions, and length of seasons are varied to obtain a 'best-fit' match between observed and simulated data.

6.2 MATERIALS AND METHODS

6.2.1 Site Description

The study site was located approximately 15 km south of the Edmonton city center at the Ellerslie Research Station. The climate of the Edmonton area is between dry and moist subhumid, characterized by relatively warm summers and cold winters. The average annual temperature is 3°C, with January the coldest month at -14°C and July the warmest month at 16°C. The mean annual precipitation is 440 mm, with 60% of it

occurring during May through August and with July having the greatest amount (85 mm). The mean annual winter precipitation (November through March) is 89 mm.

A large portion of the region, within which the study site is located, is under mixed grain cultivation, with barley being the common grain crop. The local topography is gently rolling to rolling glacial lacustrine over morainal plain. The study site is located at the toe to mid-slope position of a 1-2% west slope. Soils in the well-drained to moderately well-drained positions of the Ellerslie Research Station have been described as Eluviated Black Chernozems. Soil texture is clay to clay loam in the top meter with sand lenses occasionally occurring at depths from 0.6 m to 4.0 m. The water table at the study site generally fluctuates between 3 and 4.5 m in depth.

6.2.2 Sampling and Analysis.

Soil samples for isotope analysis were taken from a summerfallow plot and a barley plot that had standing stubble during winter. The plots were 23 m wide by 45 m long and were part of a larger study that monitored the effects of fall cultivation upon snowmelt infiltration for two winters, 1985-86 and 1986-87. Summerfallow, having no overwinter stubble, retains less snow with the consequence of less snowmelt infiltration. Under summerfallow no crops are planted and the soil is tilled to prevent weed growth; therefore the soil does not lose water to transpiration. However, evaporative loses may be higher due to lack of a plant canopy.

When possible, samples were taken to a depth of 4 m within each of the two treatments (Table 6.1). To obtain estimates of site variability replicate samples were occasionally taken to a depth of 1 m. Soil samples were taken approximately every 4 months from March 1986 to July 1987 (Table 6.1). On the first sample date (29 March 1986)

samples were taken at 15 cm intervals for the first 1.05 m and at 30 cm intervals thereafter. Samples taken during all other dates were taken at 10 cm intervals for the first 1.0 m then at 20 cm intervals after that. Samples for the top 30 cm were obtained by shovel. Samples beneath 30 cm were taken with an Oakfield sampler. Cores obtained with the sampler were scraped to remove contaminating soil particles and soil from the entire interval was bagged. To obtain sufficient sample for dry conditions in the top meter, two core holes were used. Not all samples were analyzed for isotopes.

Table 6.1. Soil water sampling for stable isotope determination.

Date	Plot treatment	Sites	Depth of samples (m)
29/3/86	standing stubble	3	1.95, 2.25, 3.75
	standing stubble	0	
22/7/86	barley	3	1.00, 1.00, 4.00
	summerfallow	3	1.00, 1.60, 4.40
04/11/86	barley stubble	1	3.60
	summerfallow	0	
14/4/87	barley stubble	2	1.00, 4.00
	summerfallow	2	1.00, 4.00
21/7/87	barley	1	4.20
	barley	1	2.20
21/10/87	barley chiselled	1	2.60
	barley stubble	1	4.00

Rain and snow samples, representative of daily precipitation events, were taken between January 1986 and October 1987 and analyzed for relative oxygen-18 and deuterium abundances. Rain samples were collected using a standard rain gauge located 0.5 km south of the University of Alberta campus. Snow samples for individual events were obtained from a raised surface to prevent accumulation of drifting 'old snow', cleaned of previous snowfall. Some snow samples were taken from accumulated snow at

Ellerslie using a snow core. As these samples are of the entire snow profile they cannot be dated to a specific event. Other relevant meteorological information were obtained from the Ellerslie meteorological station.

Water was extracted from the samples with azeotropic distillation using toluene (Allison and Hughes 1983; Hendry 1983). In this method the soil sample is suspended in a cellulose thimble above boiling toluene in a distillation apparatus. The toluene boiling at 110°C turns the soil water into vapour. Both the toluene and water vapours are condensed and accumulate in a side trap. Toluene being lighter than water and virtually immiscible floats on water. This method has an accuracy of water extraction of ±0.1 ml (Pande 1974). Residual toluene in the collected water (≈ 100 ppm) was removed with molten wax in stoppered glass test tubes.

Both deuterium and oxygen-18 were analyzed using facilities at the University of Alberta. For isotopic analysis of oxygen-18, 5 ml of water were equilibrated with CO_2 at 25°C for 24 hours followed by isotopic analysis of the CO_2 (Epstein and Mayeda 1953; Gonfiantini 1981). Analysis of deuterium involved reduction of 15 to 20 µL of sample in an evacuated tube with 0.5 g to 0.7 g of Analar zinc shot heated to 450 °C (Coleman et al. 1982). This produced hydrogen gas which was measured by a VG Micromass® 602C. Concentrations of oxygen-18 were determined using a VG Micromass® 602D. All isotope measurements are reported relative to Standard Mean Ocean Water (SMOW).

Overwinter volumetric soil moisture content was measured on the study plot with a neutron moisture meter. Water wells to depths of 3 m and 5 m were located approximately 20 m outside of the study plots.

6.3 DESCRIPTION OF SIMULATION MODEL

The model used in this study simulates one-dimensional, non-reactive, solute transfer in porous media using a finite difference equation based on rectangular integration (Leistra 1973; van Genuchten and Wierenga 1974; Bresler et al. 1982; Miller 1983). The numerical model used in this simulation may be written as (Leistra 1973):

$$C_{i,j+1} = (r_v + r_{d\,i-1})C_{i-1,j} + (1 - r_v - r_{d\,i-1} - r_{d\,i})C_{i,j} + (r_{d\,i})C_{i+1,j} \qquad [6.1]$$

where the indices i and j respectively indicate sequences in space (Δz; cm), and time (Δt; d); C refers to the solute concentration ($\delta^{18}O$ ‰); and $r_v = (q/\theta)(\Delta t/\Delta z)$ and $r_d = \Delta t((D-N)/\theta)/(\Delta z)^2$ where q is moisture flux (cm/d), D is the apparent hydrodynamic dispersion (cm^2/d), and N is the correction factor for numerical spreading. Here $r_{d\,i}$ refers to the spreading from compartment i in the forward direction. Initial conditions define C_{i,j_0} for all z. The concentration for the next time step, C_{i,j_1}, is determined from C_{i,j_0} at each depth increment and the process repeated for each time step. Thus the mass of solute transported is proportional to the volume of water moving between layers and the solute concentration in the water. This numerical solution is based upon backward differences rather than central differences. Correction for numerical dispersion is approximated by subtracting N from D (Bresler 1973):

$$N = 0.5\, q\, \Delta z \qquad [6.2]$$

with the condition that $N \leq D$. To solve equation [6.1] for stable conditions it is essential that the following conditions are held (Leistra 1973):

$$r_v + 2r_d < 1 \qquad [6.3]$$

With Eq. [6.1] a soil profile may be divided into a series of compartments in which the moisture content and hydrodynamic dispersion, may be different, but not changing, from other compartments. Flux may vary between compartments only if excess water is removed at the boundary, as with moisture withdrawal due to plant roots. Further details on model development and evaluation with a sensitivity analysis are given in the previous chapter (Chapter 5).

The input parameters for the model consisted of; length of summer and winter seasons, soil moisture content, soil moisture flux, hydrodynamic dispersion and diffusion coefficients, and $\delta^{18}O$ (‰) concentrations at the soil surface and the water table. Output was the $\delta^{18}O$ (‰) abundance for a four meter deep soil profile. The flux and flux direction was varied to obtain a match between simulated and observed $\delta^{18}O$ (‰) concentrations in the soil profile. All of these properties, except for the dispersion and diffusion coefficients, were based upon field observations.

The conditions simulated are meant to represent conditions of water table recharge during one season (summer) and upward flow to a freezing front during the other season. The upward flow conditions can also be represented by flow towards a root zone. Actively transpiring roots would result in little movement of water towards the surface. The model and the conditions simulated are simplified as much as possible in order to focus upon the processes rather than upon the model.

Initial and boundary conditions for the model are represented by the following:

C_s = surface ^{18}O ‰	at $z = 0$	for $0 \leq tn$
C_p = profile ^{18}O ‰	at $0 < z \leq L$	for $t = 0$
C_w = water table ^{18}O ‰	at $z = L + 1$	for $0 \leq tn$

q_s = downward flux (+)	at $0 \leq z \leq L$	for $0 \leq t_{2n+1}$ (odd seasons)
$q_{w1} = 0$	at $0 < z < R$	for $0 \leq t_{2n}$ (even seasons)
q_{w2} = upward flux (-)	at $R \leq z \leq L$	for $0 \leq t_{2n}$ (even seasons)

where L is the water table surface, R is the bottom of the root zone, odd seasons are summers (the first season of the simulation), and even seasons are the winters. For downward flow, solute reaching the water table (z = L+1) is considered mixed with an infinitely large reservoir of solute concentrations established for the boundary conditions (C_w). A winter 'no-flow' soil layer ($0 < z < R$) with removal of upward flowing waters at the bottom of this layer, at R, would essentially be the same as allowing downward flow from the surface and removal of water at the root zone (R) along with the upward flowing groundwater.

Equilibrium (minimum change between seasons) of solute concentrations between seasons was tested by the following expression:

$$\sum_{i=1}^{n} \left(|C_{i,d} - C_{i,d-1}| + |C_{i,u} - C_{i,u-1}| \right) \leq 0.01 \qquad [6.4]$$

where the d and u subscripts represent the solute concentration for the last time step at the end of the downward flowing and upward flowing seasons respectively. Equation [6.4] essentially compares the sum of the absolute differences between solute concentrations for all depths at the end of each season, $C_{i,d}$ and $C_{i,u}$ with that of the previous year $C_{i,d-1}$ and $C_{i,u-1}$ to check whether stability (minimum change) has been achieved. Calculations for each simulated year are repeated until t_n, when the total differences are equal to or less than 0.01.

6.4 APPLICATION OF MODEL TO FIELD DATA

6.4.1 Field Conditions.

Only two input processes are estimated, hydrodynamic dispersion using Eq. [5.3] (Chapter 5) and diffusion of water in soil using Eq. [5.2] (Chapter 5). The following is a description of field observations that helped to establish the input values and operating assumptions for the simulation model.

Winter **precipitation** and **temperature** during the study period were drier and warmer than normal (Table 6.2). Spring and fall conditions were wetter than normal, while summer precipitation was near normal. Freeze-up of the soil and development of a permanent snow cover occurred by the last week of October or the first week of November. Midwinter melts occurred during both winters, with melts during the first winter being more severe and resulting in almost total reduction of snowpack. Springmelt for the both winters began in the last half of March and finished by the first week in April with soils remaining frozen at depth usually until late April.

Table 6.2. Climatic summary of study period, Fall 1985 to Summer 1987.

	1985 to 1986		1986 to 1987		long term mean†	
Season	T (°C)	P (mm)	T (°C)	P (mm)	T (°C)	P (mm)
Sept. - Oct.‡	5.4	65.4	7.2	109.4	7.4	57.5
Nov. - Mar.	-8.6	80.8	-6.6	49.8	-9.9	89.0
Apr. - May	7.5	84.7	9.5	80.5	6.7	64.9
Jun. - Aug.	14.9	222.5	15.2	235.6	15.0	228.9

† normals are from 24 years of meteorological records for Ellerslie Research Station.
‡ inclusive.

Meteoric oxygen-18 values (‰) collected during the study period displayed a greater range in values than those of the International Atomic Energy Agency data (IAEA 1969, 1970, 1971) collected in Edmonton during 1961 to 1966 (Fig. 6.1). This is

partly due to the fact that the IAEA analyzed monthly composite samples as opposed to daily samples as done in this study; however, the summer data collected for this study also had lower and higher $\delta^{18}O$ values than daily values collected by Linton (1972) and Ellis (1979) in the Edmonton region. The wide range of $\delta^{18}O$ values in the summers of 1986 and 1987 were due to colder and warmer than normal monthly temperatures. The $\delta^{18}O$-δD water line of the 1985 to 1987 meteoric waters is calculated to be:

$$\delta D(‰) = 7.29(0.16)\ \delta^{18}O(‰) - 6.88(2.67) \qquad [6.5]$$

where; n = 52,

probability of a larger F = 0.000,

coefficient of correlation = 0.989,

standard error = 6.16

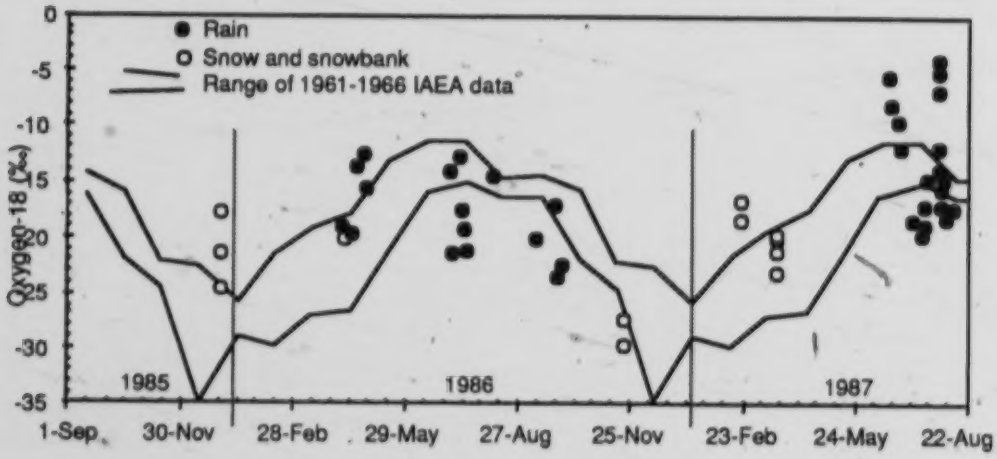


Fig. 6.1. Oxygen-18 values of meteoric water for 1985 to 1987.

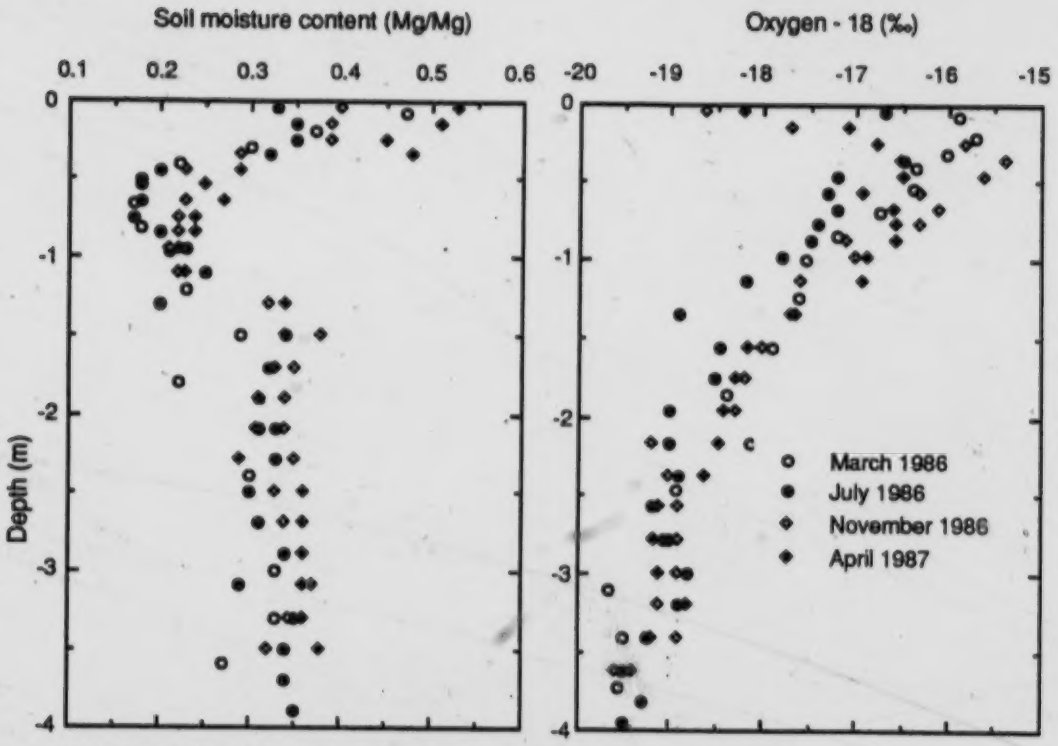
The average **soil temperature** for Ellerslie, at a depth of 0.5 m ranges from a low of -1.9 °C in February to a high of 14.4 °C in July (Phillips and Aston 1979). At 3.0 m monthly temperatures range from 2.4 to 8.1 °C. Mean monthly temperatures below 0 °C are occur at the 1.0 m depth but not at 1.5 m. The average annual soil

temperature for the depth interval from 0.2 m to 3.0 m is 5.2 °C. (Phillips and Aston 1979).

Soil porosity was estimated from bulk density (using a particle density of 2.65 Mg/m^3) and volumetric soil moisture measurements near saturated conditions, ranged from 59% for the surface 0.15 m to 39% below 0.75 m.

The standard deviation of **soil moisture content** was the highest in the top meter (Table 6.3, Fig. 6.2) due to precipitation and moisture extraction by roots, and decreased in value with depth. A sudden change in moisture content is evident at about 1.3 m (Fig. 6.2) and assuming that this is not caused by any textural changes is taken to represent the upper boundary of the capillary fringe. For most of the study period the water table varied between 3.7 m and 4.3 m below the soil surface (Fig. 6.3). Soil moisture contents in the top meter during November 1986 and April 1987 were well above field capacity (0.35 Mg/Mg ± 0.014) for the 0 to 0.3 m interval (Fig. 6.2).

Soil water isotopes decreased in both concentration and standard deviation with depth (Fig. 6.2, Table 6.3). Between sampling dates only the soil isotopic values in the top 0.3 m were significantly different (paired t-test, $p \leq 0.05$) particularly between July 1987 and the other dates. Below the depth of one meter fluctuations in isotopic values between dates can be attributed to site variability as no significant differences between dates occurred. The isotope concentrations in the soil profile are enriched by evaporation as their $\delta^{18}O$-δD values would fall to the right of the meteoric water line if plotted (i.e., if a $\delta^{18}O$ value of -15.8 ‰ were inserted into Eq [6.6] the calculated δD value would be -122.1 ‰). The deeper penetration of isotopically enriched water could be due to downward flow of soil water caused by soil water extraction by roots at and below these depths (Barnes and Allison 1983).



values in figures are not representative of all samples.

Fig. 6.2. Distribution of soil water content and $\delta^{18}O$ (‰) with depth.

Table 6.3. Isotopic composition of soil water and groundwater.

Conditions (m)	n	$\delta^{18}O$ (±sd)‰	δD (±sd)‰	θ (±sd)m^3 m^{-3}
Both plots all dates*				
0-0.3 m	27	-15.8 (2.32)	-135.2 (7.34)	0.39 (0.063)
0.3-0.6 m	27	-15.8 (0.90)	-132.4 (3.52)	0.25 (0.048)
0.6-0.9 m	22	-16.6 (0.60)	-135.4 (3.44)	0.22 (0.040)
0-0.9 m	76	-16.1 (1.55)	-134.2 (5.32)	0.29 (0.094)
1.0-2.0 m	30	-17.8 (0.90)	-143.7 (4.23)	0.27 (0.055)
2.0-3.0 m	21	-18.8 (0.38)	-149.8 (1.99)	0.32 (0.025)
3.0-4.0 m	14	-19.3 (0.26)	-151.7 (2.03)	0.34 (0.038)

* only samples which had complete analysis; oxygen-18, deuterium, and water content, were included for this summary.

Of importance to this paper is that large seasonal differences in $\delta^{18}O$ concentrations that occur for Edmonton summer and winter precipitation (Fig. 6.1) are not reflected in the soil water (Fig. 6.2). Following the 1986 springmelt, the average soil water $\delta^{18}O$ from three sites in the top 0.35 m was –15.9‰ (±1.1‰, n=9). Concurrent neutron probe measurements indicated that overwinter soil moisture increases accounted for 21% of the soil water in the top 40 cm. In July 1986, 63 mm of rain fell in seven days with an average $\delta^{18}O$ value of -19.7 (±1.6) ‰, yet only the top 10 cm of soil indicated any effect with an average isotope value of -17.4 ‰ (±0.8‰, n=4), whereas the 10 to 20 cm interval had an isotope value of -14.3 ‰ (±0.7‰, n=4).

This lack of response of the soil water isotopes to rainwater and snowmelt is likely due to a large proportion of relatively immobile soil water caused by the high clay content of the soils. The rain water mixed with the soil water rather than displacing them and likely also moved through cracks and channels to deeper depths.

During the winters of 1985-86 and 1986-87 the soil moisture at deeper depths, 85 cm tended to increase slightly, while the depth to the **water table** increased (Fig. 6.3). During the summers the depth to the water table decreased. The increase in soil moisture content can not be explained by infiltration as the soil surface was frozen and there were not concurrent increases (or decreases due to drainage) in the overlying soil. Soil water contents during the first winter were lower than the second, and the overwinter (November to March) increases at 85 cm depth were greater. It is a possibility that the drop in water table level during the winter was due to upward water flow to a freezing front. This phenomenon has been well documented by others (Ferguson et al. 1964; Willis et al. 1964; Benz et al. 1968; Gray et al. 1983). Gray et al. (1983), in Saskatchewan, measured overwinter moisture increases in the top meter of 47.8 mm for a silty clay soil with a water table below 4 m. The greater fall in

the water table during the first winter could be attributed to the colder temperatures (Table 6.2) and the drier soil moisture conditions at depths at 0.5 to 1.0 m (Fig. 6.3). Heavy rains (59 mm) at the end of July 1987 resulted in a water table rise of about 1.3 m. Ignoring effects from heavy rainfalls and taking the first winter and summer as representative of normal climatic (Table 6.2) and soil moisture conditions, the daily rate of the water table falling during the winter and rising during the remainder of the year is estimated at 4.0 mm/day and 2.9 mm/day respectively.

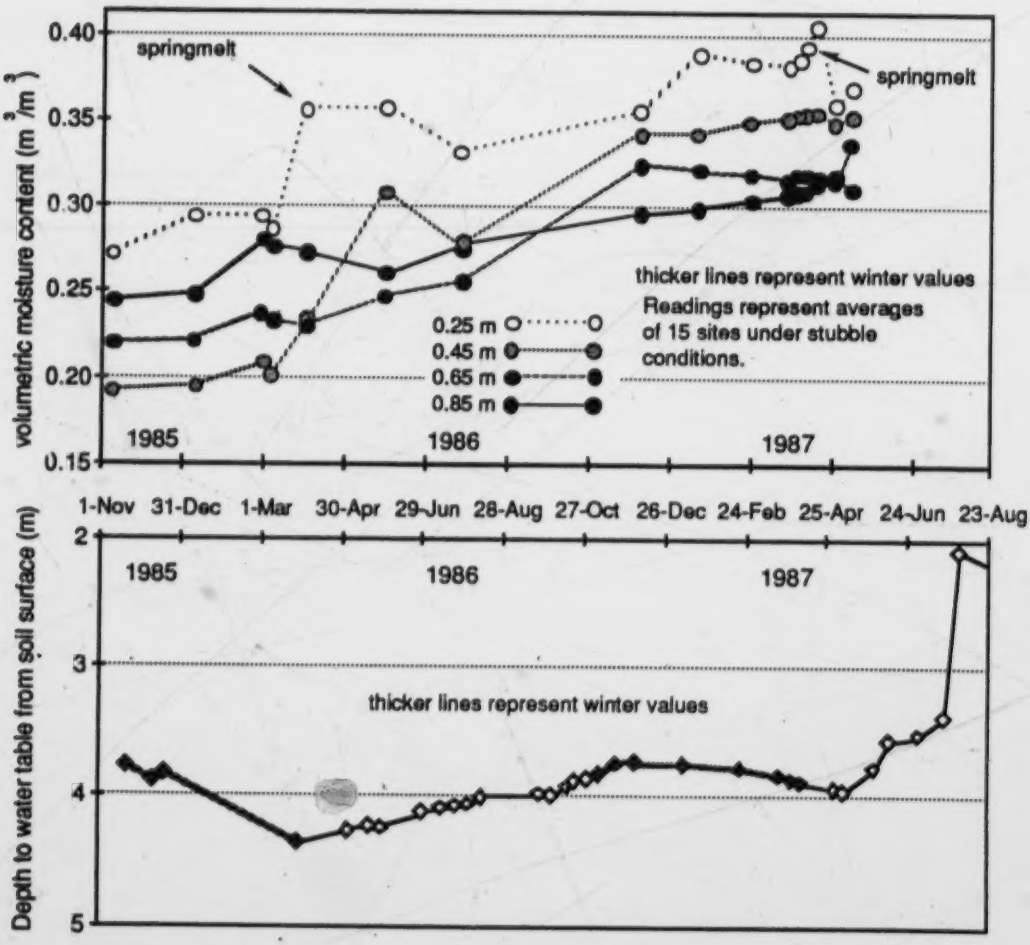


Fig. 6.3. Soil water content and depth to water table, fall 1985 to fall 1987.

Piezometric measurements taken by Sanborn (1980) at a nearby site (approximately 500 m to the west and several meters lower in surface elevation) indicated the possible upward movement of groundwater from deeper depths to a more permeable layer at 4.5 m below the soil surface. During his measurements (1978 to 1979), the head was always higher in the 10.4 m piezometer than in the 7.3 m piezometer, while the 4.5 m piezometer was always dry. A water table well (3.6 m below the soil surface) remained dry throughout the study period except for May 1979 when the water table was within 1-1.5 m of the surface. As the 4.5 m piezometer was dry this was likely a perched water table.

Overwinter soil moisture increases below 0.6 m are attributed to the contribution of capillary water to the freezing front. During the first winter there was an increase of 10 to 14 mm between 0.6 and 1.0 m, while during the second the increase was only 5 to 7 mm. Extending the depth interval of gain to the top of the capillary fringe, the total increase for the first winter is estimated to be between 18 and 25 mm. This would correspond to a change in moisture content within the region of water table rise and fall of 0.03 to 0.04 m^3/m^3 or an average upwards flow rate of 0.14 mm/day for winter (November 1 to March 31). An equivalent rise in the water table during the spring through summer corresponds to an average downwards flow rate of 0.10 mm/day. These estimates do not account for groundwater movement away from the site.

6.4.2 Model Input Values and Assumptions

The model and its input parameters are set up according to the following information and rationale and are summarized in Table 6.4:

- water infiltrates during spring through fall in the form of precipitation, and rises from the 4 m depth during winter in the form of groundwater.

- water exits at 0.6 m during spring through fall due to extraction by roots. This depth was picked due to the higher isotope values at the depths 0.3 m to 0.6 m likely caused by root extraction of soil moisture at these depths.
- the year is divided up into two seasons, winter and summer. Winter is from soil freeze up, 1 November, to soil thaw, at mid April and summer the remainder of the year, spring through fall.
- winter is represented by no water entry from the surface and no water movement in the root zone due to frozen conditions. Water rises from the water table to the freezing front located at 0.6 m.
- spring through autumn is represented by water entry from the surface, extraction by roots at 0.6 m and water recharge to the groundwater table. The average annual precipitation is 460 mm. Assuming that winter precipitation does not enter the soil until springmelt, the average daily potential (no evaporation) flow rate into the soil surface from April 15 to October 31 is 2.3 mm/day (460 mm for 199 days). The daily surface infiltration rate is set at 2.0 mm/day and the transpiration rates at 0.6 m are set at 2.0 mm/day plus or minus any contributions from or to the water table.
- the isotope concentrations of the precipitation and the groundwater entering the system remain constant throughout the modelled time period. As the isotope values of the soil profile did not change below 0.3 m during the study period, it is felt that this assumption is valid. With this assumption the effects of fractionation and enrichment by evaporation may be ignored as this is already included in the $\delta^{18}O$ profile to be modelled.
- it is assumed that the measured $\delta^{18}O$ profile (Fig. 6.2) is stable and not measurably affected by fluctuations in climate or crop. The land at the research station has been cleared and under cultivation for at least the last 80 years, with barley grown at that site for at least the last 15 to 20.
- based upon the above information and the isotope concentrations in Table 6.3 the oxygen-18 values input from the soil surface and the water table will be -15.7‰ and -19.3‰ respectively.
- Soil moisture contents are chosen from average volumetric moisture contents measured by the neutron probe. Values in the root zone are set

below field capacity, while those below the root zone (0.6 to 4.0 m) are set above field capacity. These will remain constant during the simulation time period.

- For calculation of the diffusion coefficient the average annual soil temperature of 5 °C is used.

The models were run with the parameters from Table 6.4. The flow rates between the root zone and the water table were varied between 0.01 and 0.15 cm/day. The objective of the simulation was to select seasonal flow rates between the root zone and the water table so as to match the simulated $\delta^{18}O$ soil water profile with that measured in the field. This simulation ignores any assumptions or divisions of the soil water into mobile and retained water phases.

Table 6.4. Soil, hydrologic, and isotope parameters used in simulation model.

Property	Value
Dimensions	0.0 - 4.08 m, with Δz = 0.06 m and Δt = 3 days
Root zone	0.0 - 0.60 m.
Capillary fringe zone	0.6 - 4.02 m
Seasons	Winter of 166 days, Spring to fall of 199 days
Water inflow	2.0 mm/day down from surface, spring to fall Variable rates up from water table, winter
Water outflow	Variable rates down to water table, spring to fall Out from root zone at 0.6 m
Soil properties	depth (m) porosity moisture (m3/m3) 0 - 0.60 0.45 0.30 0.60 - 4.08 0.40 0.38
Soil temperature	5 °C
Isotope input	Surface $\delta^{18}O$ of -15.7 ‰ Water table $\delta^{18}O$ of -19.3 ‰
Profile at t = 0	-17.0 ‰

6.4.3 Results of simulation run.

The flow rates estimated from soil moisture and water table measurements (summer, 0.01 cm/d; winter, -0.014 cm/d) were first entered as flux variables into the model. The resulting $\delta^{18}O$ concentrations are too high as compared to the measured values (Fig. 6.4). Lower seasonal flow rates of 0.05 and -0.10 cm/d result in $\delta^{18}O$ values that encompasses the variability of the measured values better. The model indicates that the different isotope concentrations resulting from the seasonal flows could explain of some of the $\delta^{18}O$ variability (Table 6.3).

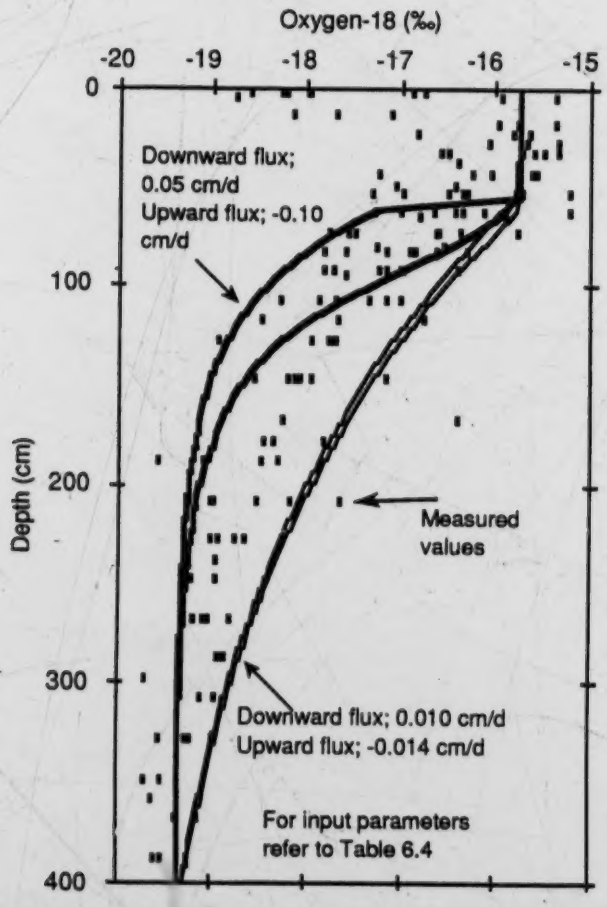


Fig. 6.4. Simulated oxygen-18 concentrations with depth for summer/winter fluxes of 0.05/-0.10 cm/d and 0.010/-0.014 cm/d.

Tested flow rates that best reproduced the profile shape for $\delta^{18}O$ had summer/winter fluxes of ratios ranging from 1:2 to 1:1.5 (Fig. 6.5). Varying the relative fluxes by 10 to 20% resulted in simulated $\delta^{18}O$ values not matching the measured values. The flow rates that resulted in the summer and winter $\delta^{18}O$ values encompassing the measured variability were 0.10/-0.15 cm/d, about an order of magnitude greater than those estimated from field measurements. As not all of the variability is due to alternating flow directions but also due to the natural variation of correlated soil properties such as porosity, texture, and moisture content, the higher flow rates are likely too high. The lower flow rates in Fig. 6.5, 0.01/-0.02 cm/d and 0.05/-0.08 cm/d are likely those which are closest to real flow values.

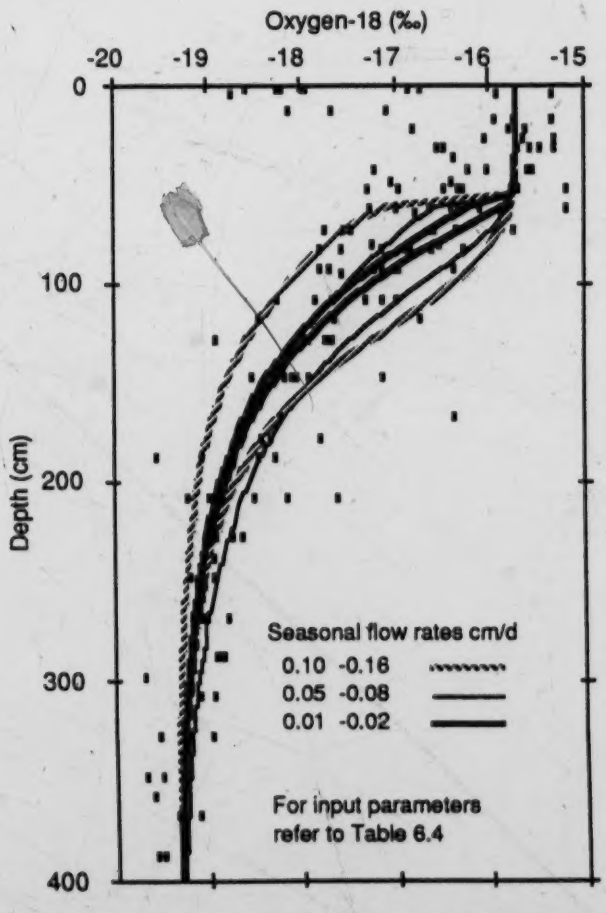


Fig. 6.5. Simulated oxygen-18 concentrations with depth for summer/winter fluxes of 0.10/-0.16 cm/d, 0.05/-0.08 cm/d, and 0.01/-0.02 cm/d.

As seasonal trends in the measured oxygen-18 values were not evident, then it is likely that the flow rates are low (less than 0.05/-0.08 cm/d) and/or variability obscures any seasonal trends. This assumes that the measured oxygen-18 values are the result of an equilibrium state caused by similar climatic and vegetative conditions during the previous 50 to 80 years. The sensitivity analysis in Chapter 5 indicated that the model is quite robust with regard to varying the dispersion coefficient and the moisture content. Varying the relative seasonal flux or the relative season length, however, results in large changes of oxygen-18 values. The accuracy of the relative flux estimates (between seasons) for the model are within 25%; however, as the relative seasonal lengths also have a strong effect, the combination of changing the season lengths and the flux could result in a multitude of correct fluxes. Other conditions can further confound the concept represented by the alternating flow seasons model: upward flux to the root zone during the summer due to dry conditions created by evapotranspiration; and water from high intensity rainstorms could flow to the water table down macropores and cracks without wetting the root zone while water could be simultaneously moving upwards in the smaller pores. As the model just considers matrix flow any rapid flow down structural cracks or old root channels due to near saturated surface conditions would bypass the mixing and flow rates used in the model. As indicated by heavy rain storms these preferential water flow routes from the surface to the water table do exist. However the model does indicate that the observed isotope changes with depth is tne result of alternating flow and that there is a greater amount of upward flowing water relative to downward movement.

6.5 SUMMARY AND CONCLUSIONS

The objective of the study presented in this paper was to recreate soil water flow conditions, using a one-dimensional flow model, that might have resulted in the observed $\delta^{18}O$ concentrations in soil profile. The measured oxygen-18 content decreased with depth between the root zone (0.3 to 0.6 m) and the water table (4.0 m) and during the period of measurement did not significantly change with time. The change in profile concentrations with depth and the stability of the oxygen-18 values with time indicate little moisture movement and the possibility of flux that alternates in direction, as flow in one direction would have resulted in uniform $\delta^{18}O$ concentrations.

Alternating water flow between the groundwater table at 4.0 m and the root zone at 0.6 m was put forward as an explanation for the shape of the $\delta^{18}O$ soil water profile. This hypothesis was based upon observations of seasonal rise and fall (summer and winter respectively) of the water table and overwinter moisture gains to the lower root zone. The hypothesis was tested with a one-dimensional finite difference model which describes the convective flow and dispersion of non-reactive solutes. Input parameters consisted of soil moisture, porosity, diffusion, and hydrodynamic dispersion. The output was $\delta^{18}O$ content with depth. Flow rates for summer and winter were varied until the simulated profile matched that of the observed $\delta^{18}O$ profile.

Matching of measured and modelled d18O profiles by numerical simulation was achieved with summer/winter flow ratios ranging between 1:2 to 1:1.5 and flows ranging from 0.01/-0.02 cm/d to 0.10/-0.15 cm/d. The higher values, 0.10/-0.15 cm/d, had summer and winter oxygen-18 profiles that 'straddled' the measured points. As variability would likely result in some scatter of isotope values about the modelled

profiles and no seasonal trends were evident in the measured values it is likely that the lower fluxes, 0.01/-0.02 cm/d and 0.05/-0.08 cm/d, best represent field conditions.

The flux values inferred from field measurements of soil moisture and water table depth, 0.01/-0.014 cm/d, resulted in over estimating the isotope values. This lack of fit could be explained by upward flow also occurring during the summer due to water removal by transpiration. Although the model supports the hypothesis that seasonal change in direction of flow occurs, there are strong indications that not all the upward flow occurs during the winter.

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7. SYNTHESIS AND CONCLUSIONS

7.1 SYNTHESIS

7.1.1 Introduction

Although the primary emphasis of the research presented in this dissertation was on soil moisture in the fall and after snowmelt for an agricultural prairie soil, the process of investigation and explanation necessitates a comprehensive consideration of the hydrologic cycle. This section provides a synthesis of the papers in this dissertation by providing a conceptual model of the field water cycle, and by evaluation of the methods of measurement used.

7.1.2 The Field Water Cycle at Ellerslie

Due to the high infiltration rates of the soils at Ellerslie (Eluviated Black Chernozem), the occurrence of runoff from maximum rainfall intensities is not expected unless surface sealing occurs (Verma and Toogood,1969). Generally runoff occurs only during springmelt. This was observed for both springs of the field program (1986 and 1987) when melt water, unable to infiltrate, ponded in small depressions. The infiltration capacity of the soil was greatly lowered due to high frozen moisture contents and surface ice layers resulting from midwinter melts during both winters of study. For example, within the disced treated plots, a cultivation that is normally used in this area, only 35.1 and 18.3% of the precipitation for the 1985-86 and 1986-87 winters, respectively, infiltrated.

Due to the low infiltration rates under frozen conditions, the amount of snow meltwter infiltrated is dependent more upon agricultural management practices and the fall soil moisture content than upon the amount of overwinter precipitation. The tillage practice which appeared to create the best conditions for trapping snow and enhancing snowmelt

infiltration was chiselled stubble for soil conditions drier than field capacity (-300 to -1000 kPa) which had an average soil moisture increase of 44 mm for the top 60 cm of the soil profile. Standing stubble had approximately 33 mm of snowmelt infiltration for dry soil conditions. Summerfallow resulted in the least amount of snowmelt infiltration (14 mm), due to the lack of surface stubble to trap and retain snow. Although wet fall conditions resulted in the reduction of overwinter soil moisture increases for all fall tillage treatments, the relative ranking of differences among treatments remained. The increase in soil moisture for the chiselled treatment was slightly greater than that for the standing stubble and both were greater than the disced treatment. The greater infiltration capacity of the chiselled plots compared to that of the standing stubble was likely due to the increased surface roughness, which trapped small pockets of snow and runoff, and the creation of large macropores that did not become clogged by frozen water as easily as those in the untilled, standing stubble.

The water table was observed to fall during the winter and rise during the remainder of the year, with inflection points occurring in late fall and mid-spring. The rise of the water table the second spring (1987) was greater, likely due to the combination of soil moisture being near field capacity and higher than normal spring rains. Only once during the study was a sudden rise in the water table measured: 59 mm of rain in the last days of July 1987 resulted in a water table rise of approximately 1.3 m. Because of the very high rainfall intensity of this one storm, part of this rise could have been the result of air pressure.

Overwinter soil water contents at 85 cm tended to slightly increase greater than could be explained from drainage. It is an unconfirmed possibility that this could be due to upward water movement towards the freezing front. The overwinter drop in the water

table could be indicative of this, but cannot be validated without further investigation into the groundwater regime of the site.

The use of environmental isotopes, deuterium and oxygen-18, helped to extend the interpretation of hydrologic processes at Ellerslie. Other studies have used seasonal variations in meteoric isotopes to trace the wetting front of a snowmelt event up to two years (Saxena 1984). This was possible for the sandy soils of Saxena's study as infiltrating water in such soils tend to displace, rather than mix with, resident water. At Ellerslie high clay content and aggregates in the surface layer provide a high proportion of relatively immobile or inaccessible water, while structural cracks provide preferential flow routes for near saturated surface conditions. This mixing of infiltrating water with soil water, rather than displacement, was indicated by the fact that soil water isotope values below a depth of 30 cm failed to reflect the large fluctuations of isotope concentrations in meteoric water.

Groundwater, at a depth of 4 m, had a lower concentration of heavy isotope than soil water (after accounting for the effects of evaporation), indicating that they were composed of a greater amount of winter precipitation (44% as opposed to the 11 to 19% of soil water). As indicated by $\delta^{18}O$-δD relationship, groundwater was not as greatly affected by evaporation as soil water was. Groundwater was estimated to be composed of 44% soil water, the remainder was precipitation that was composed of 62% winter precipitation. The precipitation portion was unaffected by evaporation as it either drained through preferential flow routes too fast to be altered by evaporation, or surface evaporation rates were too low to cause noticeable isotopic enrichment.

A simple conceptual hydrological model is presented as an explanation for the observed changes in overwinter soil moisture, changes in water table levels, and the different isotope compositions of the soil water and groundwater (Fig. 7.1). Low spring infiltration rates, due to ice filling the macropores and cracks, results in large amounts of snowmelt runoff accumulating in small local depressions. Groundwater mounding and lateral flow then occurs to the surrounding area. Some of the rise in the the water table level observed during spring and early summer could be due to contributions from lateral flow. It is assumed that intermediate or regional flow regimes have no significant effect upon these seasonal flow dynamics.

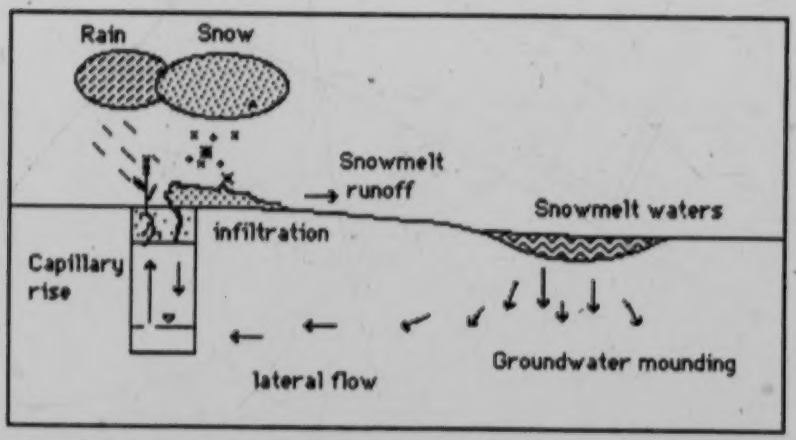

Fig. 7.1 Conceptual representation of the hydrologic cycle at Ellerslie.

Dry soils and high summer infiltration rates result in very little summer runoff. High evapotranspiration rates likely prevent water from leaching to the water table for all but the most intense rain showers. Isotopic analysis indicates that there is some contribution of root zone water (44%) to the groundwater, however, simulation

analysis of the profile shape for isotope concentration indicates that flow is not dominantly downward.

Although evaporation of soil water usually takes place in the top 10 cm, the soil water to a depth of 0.6 m was isotopically enriched due to evaporation. This indicates that there is movement from the soil surface to the root zone, likely due to a matric potential gradient resulting from crop transpiration. This is confirmed by the fact that the isotope concentrations decreased in value below 0.6 m, indicating a lack of dominant downward movement and mixing with upward moving water. It would thus be expected that other solutes (nutrients, salts, and pollutants) in the soil water, would also follow the same pattern.

7.1.3 Seasonal soil moisture flow

Inherent in the conceptual model represented in Fig. 7.1 is seasonally alternating flow between the soil and groundwater regimes. In the non-depressional areas there was a lack of dominant downward movement of the soil matrix water. This was indicated by three observations; the soil water and the groundwater were isotopically different; they did not change in value during the duration of the study; and most of the isotopic change in the profile occurred in the top 1.5 m. Isotope concentrations decreased from their highest value at 0.3 to 0.6 m (a $\delta^{18}O$ of -15.8 ‰) to an almost constant value by 2 m (-18.8 ‰).

The hypothesis that the isotope concentration profile is the result of seasonal flows that alternate in direction was tested with a convective-flow dispersion model. If the profile is assumed to be the result of long-term stable conditions (greater than 20 years) of climate and vegetation, then one-directional flows will not produce such a profile.

Fields in the study area have been cultivated for at least the last 70 years, and in barley for the last two decades and possibly twice that long. It is thus possible that the isotope profile represents a steady state situation. Reproduction of the measured isotope profile with the numerical model could only be accomplished with flow that seasonally alternates in direction. The seasons were represented by frozen and non-frozen conditions, with upward flow occurring to a freezing front. Seasonal length was set to that of Edmonton winter and spring through fall conditions (166 and 199 days respectively). The simulated flux rates that best matched simulated soil water $\delta^{18}O$ with measured were less than 0.08 cm/d for summer and winter conditions, with the upward fluxes 1.5 to 2 times larger than the downward fluxes.

The simulation does not account for fast, downward flowing water in preferential routes or lateral recharge from spring mounding in nearby depressional areas. If upward flow occurs due to a potential created by freezing, then upward flow could also be expected to occur due to the drying of the root zone by high evapotranspiration rates. Downward flow to the water table would most likely occur during spring thaw from melt water, during spring and fall when evapotranspiration is low, and during late June to early August when rain storms have the highest intensities.

The lack of dominant downward movement suggests that there would also be little downward movement of nutrients, salts or any pollutants out of the root zone. This coupled with the upward flow, that is 1.5 to 2 times greater than the downward flow, would tend to keep nutrients near the surface. The potential for downward movement of solutes would be greatest during extended wet conditions in which the inaccessible soil water would have greater time to equilibrate with the more mobile soil water through

diffusion. A heavy rain or rapid melt event could then be more effective in removing nutrients.

7.1.4 Use of environmental isotopes in soil water studies

The course of investigation into the contribution of snowmelt water to the soil and groundwater regimes has provided insight into the hydrologic cycle at Ellerslie; however, any explanation afforded by scientific research must be considered in terms of the methodology and techniques used in the study. The main tool of investigation used in this study was stable isotopes present in meteoric water.

Original project plans considered using the large seasonal differences in meteoric isotope concentrations as a means of tracing the snow melt wetting front in the soil for months after the occurrence of spring snowmelt. Several factors thwarted this plan: the relatively low amounts of winter precipitation and snowmelt infiltration; a large amount of relatively inaccessible soil water due to the high clay content; the presence of preferential flow routes (i.e., root channels, drying cracks, structural cracks, faunal activity) which would result in rapid movement of water from near saturated surface conditions to deeper portions of the soil; and abnormal weather conditions which resulted in lower than normal isotope concentrations in summer rains and likely higher than normal concentrations in winter precipitation. Possibly the factors which contributed most towards the lack of a discernible snowmelt front are the high clay content and preferential flow routes which resulted in the mixing of incoming water with the resident water rather than a displacement of the resident water.

Knowledge of isotopic patterns and variations in meteoric water is vital for interpretation of concentrations in soil water and groundwater. A study of Edmonton

meteoric water revealed that isotopic abundances and the meteoric water line for 1986 to 1987 were different and more variable than previous years of measurement (1961-66, 1966-69, 1970, 1974, 1975, and 1976). The meteoric water line had shifted slightly from that measured in 1961 to 1966. The shift, consistent for winter and summer precipitation, could be interpreted as that due to drier, more inland atmospheric water that is from re-evaporated precipitation rather than directly from ocean atmospheric waters. The two summers of measurement, 1986 and 1987, had a greater range in isotope values than any other summer. This diminished the ability to differentiate seasonal waters in the soil profile. Future use of isotopes in hydrological studies should consider the following:

- the isotope concentrations of immobile and mobile waters may be different especially if near saturated conditions result in preferential flow to the water table. This would provide information about mixing and residence time of water and possibly about groundwater contribution.

- the isotope content of snow may become enriched as it ages and exchanges with water vapour in the air. Freeman (1986) found some evidence that this occurred and that the enriched snows, and meltwater, still remained on the meteoric water line. This needs to be investigated further.

- enriched isotope concentrations due to evaporation were found in the root zone. The presence of this water is likely due to downward movement caused by diffusion and/or convection as a result of water extraction by roots. Soil water movement to roots is not supposed to involve fractionation of isotopes; however, Allison and Hughes (1983) found evidence that for very dry conditions roots shrink and vapour movement near the roots and in large macropores could have been the cause for the isotope enrichments they observed at depth.

- with concerns being expressed about changing weather patterns and global warming trends, it is possible that isotope abundances in meteoric water could be affected, especially for individual storms.

- greater accuracy in prediction equations of isotope abundances in meteoric water could be achieved if vapour origins and histories were considered.

7.2 CONCLUSIONS

Four general conclusions are drawn as related to the objectives presented in the introduction:

1. *To determine the effect of fall cultivation practices upon snowmelt infiltration.* Summerfallowed soils present the smoothest surface and consequently trap the least snow and have the least amount of snowmelt recharge. Fall chiselling is more effective than standing stubble for snowmelt recharge for a wide range of soil moisture conditions. Discing is intermedate between chiselling and fallow in recharge efficiency: much of the stubble was incorporated into the soil surface by discing thus presenting little impediment to trap snow. However, discing has a slightly better recharge efficiency than the stubble plots. This could be due to the disturbed 15 to 20 cm surface layer with a higher amount of macroporosity unlikely to become filled with frozen soil moisture.

2. *To 'typify' the isotopic abundances and fluctuations of Edmonton meteoric water.* An average difference between winter (November through March) and summer (June though August) oxygen-18 concentrations of 12 ‰ can be expected. However, if only those rainfall amounts likely to infiltrate (> 5 mm/d) are considered, this difference would be reduced by about 1.5 ‰. Single precipitation events within any season can vary by a range of 6 to 8 ‰. Variation between seasons can also be extreme especially if the two years of study, 1986 to 1987 were used as examples. Using air temperature, $\delta^{18}O$ values may be predicted to within a standard error of between 2 and 3 ‰. Origins and histories of atmospheric water vapour before the precipitation

event appear to be the main contributor to this unexplained variation both within and between seasons.

3. *To determine the relative composition of soil water and groundwater in terms of snow and rain.* Isotope analysis enabled an estimate of snowmelt contribution to soil water and groundwater, and an estimate of the contribution of water unaltered by evaporation to the groundwater. This could be interpreted as 'fast leaching' water. Groundwater, as compared to soil water, is composed of a greater amount of winter precipitation and are less affected by evaporation. This indicates that groundwater is different than soil water in both origin and history. Findings in this section tend to raise more questions about the dynamics of the unsaturated-saturated soil water systems than answers.

4. *To determine the degree of interaction and the seasonal flow rates between the soil and groundwater systems.* Analysis of the soil $\delta^{18}O$ profile using finite difference modelling methods supports the hypothesis that moisture flow between the root zone and the water table alternates seasonally in direction. The rate of flow is less than 0.1 cm/d and likely less than 0.05 cm/d. Upwards flow exceeds that of downwards by 1.5 to 2 times. Some upward flow likely occurs during the summer in response to a matric potential gradient created by a large evapotranspiration demand by the barley roots at 0.3 to 0.6 m. As the findings in this section were suggested only by two winters of soil water measurements and as the model is for one-dimensional homogenous flow conditions, caution must be taken with extrapolation of these findings.

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